

The seasonal upwellings in the Gulf of Guinea

JEAN-MARC VERSTRAETE

ORSTOM, Institut Océanographique, 195 rue Saint-Jacques, 75005, Paris, France

Abstract – Observations as well as numerical experiments show that Ekman divergence neither at the equator nor at the coast can explain a significant part of the annual upwellings in the Gulf of Guinea. The remote forcing theory is tested by analyzing both historical data and the FOCAL hydrographic stations (1982-84). The meteorological data base is analysed as well.

Analyses of the thermostad show that waters with the same characteristics as those of the South Atlantic Central Water (SACW) are carried by the equatorial under current (EUC) and the South Equatorial Undercurrent (SEUC) from the western-central equatorial Atlantic at 35° - 29°W towards the thermostad regions in the Gulf of Guinea.

The thermostad water is then carried back westward in the two geostrophic compensation flows flanking the EUC at 2° - 3°N and 2° - 3°S.

The cooling of the subthermocline waters off Abidjan (5°N) and Pointe-Noire (5°S) is related to the advection and vertical spreading of SACW between 500m and 50m in May-June-July.

An exceptional warm event occurred in early 1984 in the Gulf of Guinea. The winds collapsed nearly completely all along the equator in January-April 1984. The onset of the African monsoon was late (in July-August 1984), then the monsoon was weaker than climatology in August-October 1984. The equatorial thermocline was observed 50m deeper than climatology in February 1984, then near the surface in July 1984. Sea levels rose over the equatorial Atlantic, the surge reaching about 13cm at 6°E in February 1984, when the western-central equatorial Atlantic was nearly flat. Minimum sea levels occurred in June-July 1984 at 6°E, before the African monsoon.

During this event, the maximum speed in the shallow core of the equatorial undercurrent at 4°W and 6°E was nearly insensitive to large changes in the thermal structure (downwelling or upwelling situations). In February-May 1984, we have observed distinct increasing eastward flows at the equator below 250m and minimum equatorial thermostad thickness. In May 1984, a deep eastward jet was observed at 4°W carrying about $3 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ within one second mode Rossby radius of deformation and below the base of the thermostad, clearly separated from the EUC. Then, maximum thermostad development was found in July 1984, related to the shoaling of the deep jet and of the EUC. The top of the deep jet had shoaled to about 200m and its transport increased to about $4.6 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ within 1°30'N - 1°30'S. The spreading of the isotherms from about 300m is indicative of a geostrophic balance. Simultaneously, the equatorial thermocline was uplifted near the surface, although the base of the thermostad (13°C isotherm) remained nearly stationary.

Analysis of the perturbation temperature field shows that variations of the 19°C isotherm depth as well as the thickness of the equatorial thermostad were strongly equatorially trapped, with scales associated with the second baroclinic mode.

In the absence of local forcing in the Gulf of Guinea from January to July 1984, the only causal effect to explain these large perturbations in the upper 400m lies in the changes in the zonal wind stress to the west of the Gulf.

The distinct semi-annual cycle observed in the thermal and salinity structures in the Gulf could be attributable to the semiannual signal in the zonal wind stress to the east of 30°W. Numerical experiments by PHILANDER and PACANOWSKI (1986a) confirm that remote forcing by changes in zonal wind stress in the western-central equatorial Atlantic is the main process in seasonal changes in the thermocline depth in the eastern equatorial Atlantic, and a secondary process along the coasts.

The thinning of the thermostad in July 1984 at 3°-4°N associated with a strong Guinea current, concomitant with its thickening at the equator associated with a deep eastward jet, suggests that the upwelling along the coast is essentially a consequence of the equatorial adjustment in response to the zonal wind stress in the equatorial zone 10°W - 35°W.

Continuous observations of wind stress at St Peter and St Paul rocks (1979-1988) indicate that the trade winds relax and strengthen through a series of strong bursts superimposed on the seasonal cycle, and that an impulse forcing is appropriate from intraseasonal to seasonal time scales.

Since the equatorially trapped waves theory is now able to explain many aspects of both the 1984 winter warm event and the annual cycle, it is very likely that the remotely-forced equatorial dynamics govern not only the annual cycle but also the "quasi El Niño" in the eastern equatorial Atlantic Ocean.

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1. INTRODUCTION

The equatorial and coastal upwellings observed each year during the Northern hemisphere summer in the eastern equatorial Atlantic ocean exhibit perplexing features. The forcing mechanism responsible for these upwelling events, associated with low sea surface temperature (SST), has been a subject of speculation and considerable research. Until now, there has been no

convincing demonstration of a direct relationship between the local winds and the cold SST, either at the equator or along the coasts (VERSTRAETE, 1970a,b; HOUGHTON, 1976; BAKUN, 1978; VOITURIEZ, 1981a,b; PICAUT, 1983). Detailed analysis of the SST and subsurface thermal structure by HASTENRATH and LAMB (1977), MERLE (1978), and HOUGHTON (1983) reveal that the equatorial and coastal regions are separate.

The surface coolings are not the result of the seasonal variations of the heat gain from the atmosphere through the surface and are not confined to the upper mixed layer (MERLE, 1980; VERSTRAETE, 1985a,b, 1987). Cold waters can be advected along the northern coast by the westward Guinea undercurrent observed by LEMASSON and RÉBERT (1973a,b), and also along the southern coast by the northward Benguela current observed south of about 10° - 15° S (LONGHURST, 1962). However, the Guinea undercurrent is too small to account for the upwelling by advection of cold water from the eastern Guinea Gulf (PICAUT, 1983), and GORDON and BOSLEY (1991) show that the southward flowing Angola current along the African coast creates, with the Benguela current, the Angola-Benguela front at about 15° S (SHANNON *et al.*, 1987), so that it is very unlikely that the cold waters observed in July and August along the southern Guinea Gulf coast originate from the Benguela current. On the other hand, VOITURIEZ (1981b) found no seasonal increase of vertical mixing processes at the equator during the CIPREA experiment in 1978-1980. Recent studies by GOURIOU (1990) confirm this result and give evidence of a strong mixing in July-August, attributable to higher shears between the South Equatorial Current (SEC) and the Equatorial Under Current (EUC). However, the surface cooling associated with the higher shears and vertical turbulent mixing was estimated to be 1.6°C (between 35° W and Africa, within 1.5° of the equator) which is only about one third of the seasonal cooling observed to the east of 20° W.

If none of these possible mechanisms (vertical mixing, advection, heat gain from the atmosphere) nor the local winds are responsible for the upwellings in the Gulf of Guinea, what then is? Only studies of the response of the ocean to either impulses or low frequency basin-wide scale winds can give an answer to this question. This work presents new analyses of the observations raising from the FOCAL/SEQUENTIAL programmes during 1982-84 and the present TOGA program. It is possible from these analyses to provide a qualitative evaluation of the remote forcing mechanism responsible for some of the annual and interannual changes in the currents and the thermal structure in the Gulf of Guinea. The lack of synoptic sea level observations along the North-South boundary of the Gulf prevents us from correlating the equatorial sea level data with the sea level data along the coasts. However, a careful study of the upper water masses and the thermocline bolsters the hypothesis of a probable connection between the equatorial upwelling and the coastal upwellings in the Guinea Gulf. This study is divided in five parts organized as follows: Firstly, I give a brief review of the historical atmospheric data at the sea level over the equatorial Atlantic (pressure and winds), the hydrographic data and the synoptical data with their processings. Part two is a review of the different processes whereby wind forces the ocean with a brief summary of the equatorial wave theory. In Part three, the water masses and their mean seasonal cycle are analyzed, and the thermocline and the dynamic topography during 1982-84 are examined within the framework of the equatorial dynamics. Part four analyzes the downwelling warm event of February 1984 and the upwelling at the equator in 1984, and gives new evidence for an equatorially trapped phenomenon which bolsters the existence of a second baroclinic mode in the equatorial Atlantic. These results are discussed in Part five before the conclusion, and a coherent scheme of the equatorial dynamics in the Guinea Gulf is presented.

2. THE DATA

2.1 The atmospheric sea level pressure and the wind stress over the equatorial Atlantic

The seasonal cycle of the sea level barometric pressure over the tropical Atlantic is under the control of the north and south anticyclones (referred to the Azores and St Helena subtropical highs). According to the (1948-1972) time series archived by BUNKER and GOLDSMITH (1979), high pressure zones straddle the middle of the ocean "channel" between Africa and South America, and low pressures occur along the continents. Along the equatorial band between 10°N - 10°S is a low pressure zone (Fig. 1a). The synchrony of the annual cycle at St Helena and between the cycle observed at selected stations in the Gulf of Guinea (Fig. 1b) indicates the strong influence of the southern anticyclone over the Gulf. All over the area, the yearly variation is of about 5hpa (or mbar), with maxima in July-August and minima in February-March (VERSTRAETE, PICAUT and MORLIERE, 1980; VERSTRAETE, 1988). Other studies have shown that the barometric pressure is nearly in equilibrium with the mean sea level at the seasonal time scale in the Gulf (PICAUT and VERSTRAETE, 1976).

The wind stress over the equatorial Atlantic is available from the data files and the works of HASTENRATH and LAMB (1977), HELLERMAN (1979), HELLERMAN and ROSENSTEIN (1973) and TOURRE and CHAVY (1986).

In 1982-84, ship observations of monthly pseudo wind stress V^2 were provided by SERVAIN, SEVA, LUKAS and ROUGIER (1987) over the tropical Atlantic. In this study, I have used a constant drag coefficient of 1.2×10^{-3} to compute the wind stress in 1982-84 (LARGE and POND, 1981). For each of the eight FOCAL cruises, the zonal wind stress τ_x in the equatorial strip 2°N - 2°S was averaged every 2° between 40°W and 10°E . The meridional wind stress τ_y was computed every 2° from 6°S to 6°N along 2° wide meridional bands at 3°W , 1°E and 7°E in the Gulf of Guinea. The ship board winds observations are in good agreement with other observations. For example, the onset of the trade winds in the western equatorial Atlantic (30° - 40°W) was seen to occur in April-May in 1983 and in May-June in 1984 both in data from the meteorological station at St Peter and St Paul rocks RSPP (KATZ, 1987a) and in shipboard observations. HOUGHTON (1989) comparing data from RSPP and 4°W has demonstrated that the ship winds are excellent estimates of the real winds.

West of 0° - 5°W , the easterly trade winds prevail. The intensification of the westward wind stress first begins in April between 0° and 10°W and then moves westward. Between 0° and 30°W , the zonal component of the windstress presents a biannual signal, with two maxima and two minima. At 30° - 40°W , winds are weakest in April and strongest in September, increasing by about 0.5 dyne cm^{-2} (Fig. 2a).

East of 0° - 5°W , the southerly winds veer southwesterly into the Gulf of Guinea and become westerly off the African coast. At 2° - 6°N , the alongshore winds are weakest in May and strongest in August and they increase by less than $0.25 \text{ dyne cm}^{-2}$ (Fig. 2b). The northward wind stress intensify everywhere in the equatorial band 2°S - 2°N , between April and August-September (Fig. 2c), with a maximum increase at 30° - 40°W ($0.85 \text{ dyne cm}^{-2}$) and a minimum increase to the East of 0°W ($0.20 \text{ dyne cm}^{-2}$). This annual intensification of the southerly winds in the Gulf of Guinea is also called the African monsoon. In summary, east of 0° - 5°W and north of 10°S , the seasonal variability of the alongshore winds along the coasts of the Gulf of Guinea is very weak everywhere.

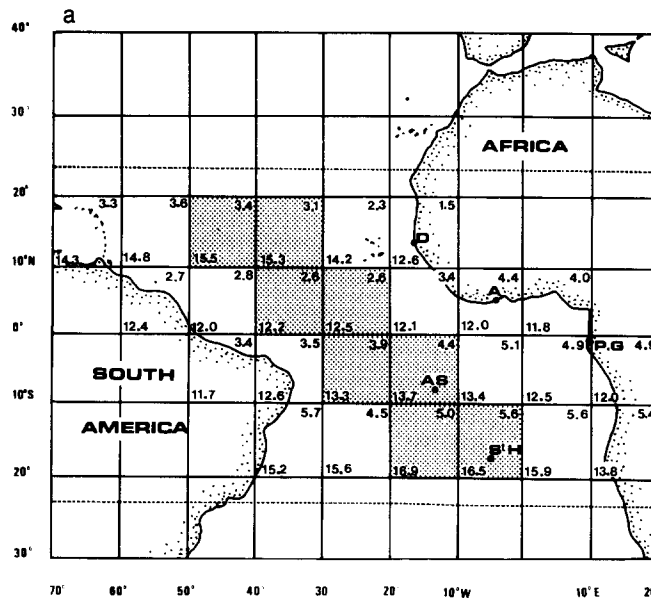


FIG.1a Annual amplitudes from crest to trough of the sea level pressure over the Tropical Atlantic (1948-1972). Upper number in each square is amplitude in hPa (mbar); lower number is climatic mean pressure (-1000) in hPa. The region of a maximum sea surface pressure area is dotted. SH: S. Helena; AS: Ascension; D: Dakar; A: Abidjan; PG: Port Gentil. (After BUNKER and GOLDSMITH, 1979).

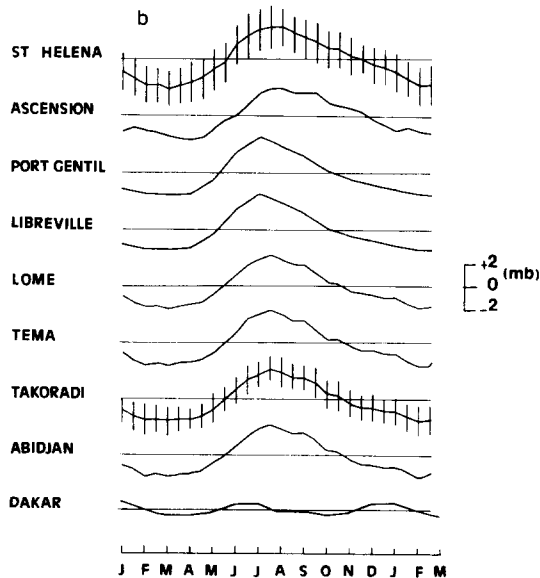


FIG.1b. Seasonal cycle of the sea level pressure at selected stations in the central-eastern Tropical Atlantic. Vertical bars show one standard deviation. Data sources: British Meteorological Office, ASECNA, Ghanaian Meteorological Survey.

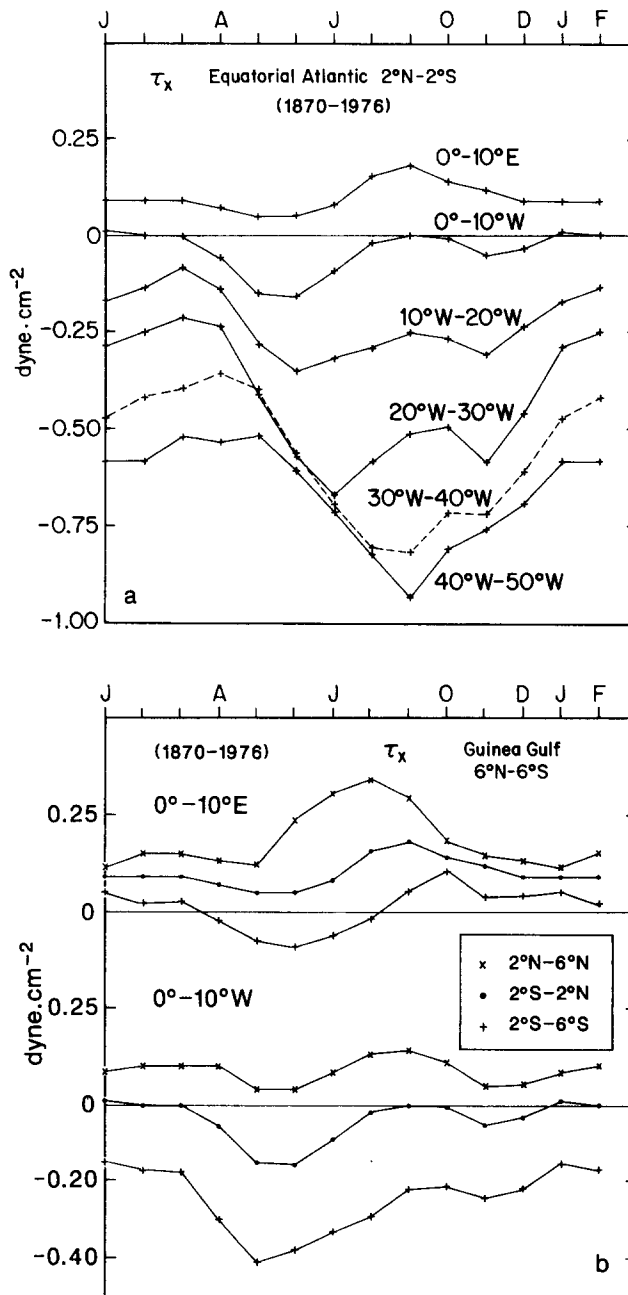
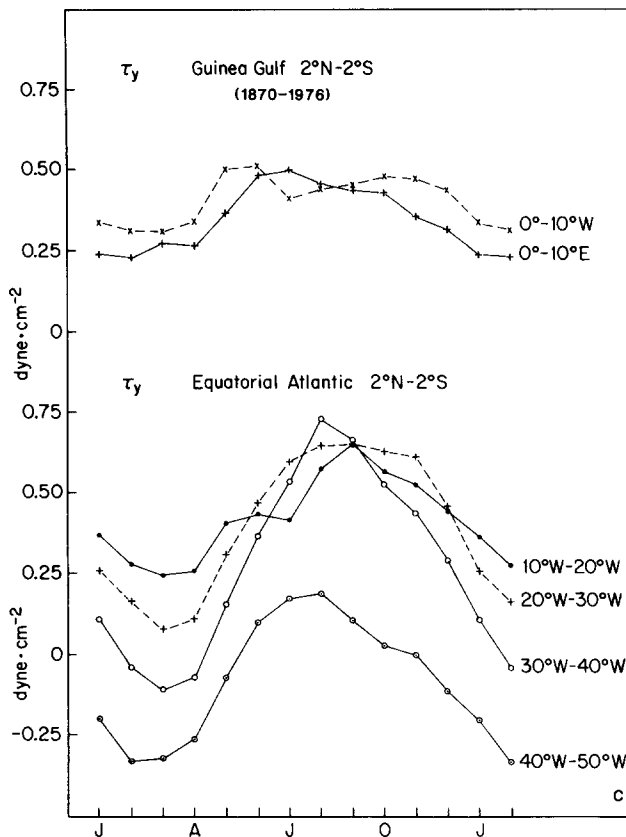


FIG.2. (above and opposite) Annual cycle of the zonal (τ_x) and meridional (τ_y) wind stress (1870-1976, all after HELLERMAN, 1979). (a) τ_x along the equator between 10°E and 50°W; (b) τ_x between 6°N and 6°S in the Gulf; (c) τ_y along the equator between 10°E and 50°W.



2.2 The historical and the FOCAL hydrographic data

The historical Nansen bottle data on file at the IFREMER Centre de Brest have been analyzed in $2^\circ \times 2^\circ$ selected areas in western and eastern tropical Atlantic and along the equator at 29°W , 10°W , 4°W and 6°E . In each area, temperature and salinity were collected at standard levels between 0 and 500m. These data are of variable quality, so for each month mean profile and temperature-salinity (T/S) diagrams have been plotted and the standard deviations computed. Any plotted values which exceeded two standard deviations from the mean have been suppressed and replaced by linearly interpolated values. Finally, the monthly statistics were computed at each level after corrections and the heat contents and the steric heights estimated. A very important set of hydrographic data was collected by the ORSTOM centres in Ivory Coast and Congo and by the Fishery Research Unit in Ghana. As a result of the observations made at these centres, the seasonal cycle is very well documented on the Ivoirian, Ghanaian and Congo-Gabon shelves (MORLIERE, 1970; VERSTRAETE, 1970b; MORLIERE and RÉBERT, 1972; WACONGNE, 1988). We have collected the data from the Nansen casts collected during 22 cruises on the Ivoirian shelf between 2°W and 8°W (Kennedy cruises, July 1969 - January 1972), and computed the dynamic heights along the 25m, 65m and 200m isobaths. The “Abidjan-Grand-Bassam” section at 4°W was observed over five years (1966-1970).

In 1982-84, 1087 stations were occupied during eight FOCAL campaigns from two research vessels using either a Neil Brown Instrumental systems CTD (conductivity-temperature-depth) between 35° W and 4°W or Nansen bottles along the transects at 1°E and 6°E. Casts were made to 500dbar. The CTD data were processed to 1 dbar resolution and gridded to 10m. The Nansen data were linearly interpolated between vertically adjacent bottles and also gridded to 10m. Systematic analysis of the T/S plots revealed that the CTD salinities were generally underestimated after the second cruise. GOURIOU, REVERDIN and DU PENHOAT (1987) established the necessary corrections for cruises 3-8 and the dynamic heights have been recomputed, using the corrected data and the most recent equation of state of seawater (UNESCO, 1981). This hydrographic data set (1982-84) has not been combined with the historical data. Moreover, the large warm event which occurred in 1984 in the equatorial Atlantic has been well documented with synoptical observations (moorings, tide-gauges and inverted echo sounders) which provide a better contrast with seasonal cycles observed from the historical data. We have selected four meridional and one equatorial sections to describe the mean temperature and salinity fields during 1982-84 (Fig.3): the 35°W section is most representative of the western equatorial Atlantic, whereas the 4°W, 1°E and 6°E sections are typical of the Guinea Gulf. For each section the mean temperature, salinity and dynamic heights with their standard-deviation were computed; at 35°W at 28 levels (0-200m by 10m step, then 240, 280, 320, 360, 400, 450, and 500m); and for the three other sections at 23 levels (0-120m by 10m step, then 140, 160, 180, 200, 240, 280, 320, 360, 400, 450, 500m). There were 21 stations at 35°W and 4°W (5°S - 5°N); 23 stations at 1°E (6°S - 5°40'N) and 17 stations at 6°E (5°30'S - 3°30'N) giving a latitudinal grid of about 1/2° everywhere.

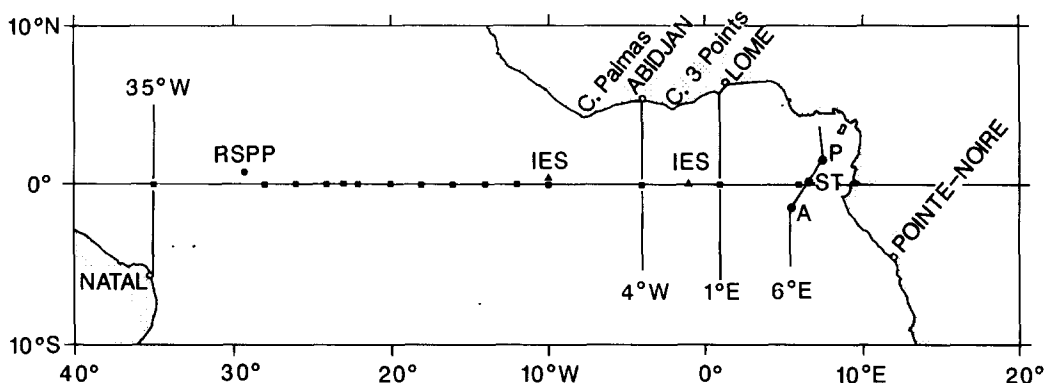


FIG.3. The "FOCAL" sections (0-500m) at 35°W, 4°W, 1°E and 6°E, October 1982 - August 1984. 1087 stations were occupied during 8 campaigns. Pressure tide-gauges were deployed by ORSTOM at St Peter and St Paul rocks (RSPP), Principe (P), Sao Tomé (ST) and Annobom (A). Two Inverted Echo Sounders (IES) were deployed by Lamont at 0°, 10°W and 0°, 1°W. The locations of the 15 equatorial stations during FOCAL are also shown.

Eight sections of direct PCM (Profiling Current Method) measurements from October 1982 to August 1984, along 4°W (6°E) and between 5°N and 5°S (1°30'N - 3°30'S), offer eight quasi-synoptical observations of the currents relatively to 500m, in the upper ocean 0-500m. Comparisons with absolute measurements at five levels in the upper 110m from one mooring at 0° - 4°W, show that a profiler drifting freely below a surface buoy gives reasonable relative currents (HISARD and HÉNIN, 1987). According to GOURIOU (1990), the PCM observations are reliable

below 15m but underestimate the absolute zonal current, particularly in the core of the EUC (18 cm s^{-1} at 4°W). Although we have no comparisons below 120m at 4°W , GOURIOU (1990) shows that at 28°W and 24°W , the differences (absolute current - PCM current) are as small as 2.5 cm s^{-1} at 200m and near zero at 300m. We estimate that, below the core of the EUC, the errors on the relative PCM currents from a buoy are less than 10 cm s^{-1} . At 6°E , the PCM was used from a ship, and the errors are probably more important but difficult to appreciate in the absence of one equatorial mooring. The meridional sections of the mean zonal currents were constructed after a careful check of each current profile, at 4°W (0-500m) and $6^\circ 30'\text{E}$ (0-400m).

The mean dynamic topography of temperature and salinity along the equator have been computed from 15 equatorial stations between 35°W and 6°E (Fig.3).

2.3 The synoptical tide-gauge and Inverted Echo Sounder data

The sea level data base used for this study was based on an array of tide gauges sited at four islands, at or near the equator, and five harbours along the coast of western Africa. The tide gauges were of two types: five were temperature-pressure (T/P) recorders which were set 5-12m depth (at St Peter and St Paul rocks (RSPP), Principe, Sao Tomé, Annobom islands and inside the Lomé harbour), and four were classical tide gauges (with float and sitting well), serviced by the hydrographic services at Pointe-Noire (Congo), Takoradi (Ghana), Abidjan and San Pedro (Ivory Coast) harbours. These tide gauge data were collected over a seven-year period beginning in January 1982. A full description of the cradles of the T/P recorders and the techniques of deployment by a diver were given by LABARRE (1983), and details of their positions, depth, time origin and nominal data records until December 1984 by VERSTRAETE (1988). In the Gulf of Guinea, there were seven recovery-redeployment operations from our small research vessel *A. Nizery* (ORSTOM) during the FOCAL program, and during the TOGA program there were five oceanographic campaigns scheduled to start from Lomé and end at Pointe-Noire between 1985 and December 1988 (Fig.4).

Data from two Inverted Echo Sounders (IES) collected at 0° , 10°W and 0° , 1°W over a three-year period between 1983 and 1987 were generously given by E. Katz of Lamont Observatory (Columbia University) who described their processing (KATZ, 1987a).

The duration, start time and observational periods differ for each record (Fig.5) so that over the seven years of active measurements in the Gulf of Guinea, relatively few observations by the different instruments at different locations overlap. Moreover, some records contained large gaps which are impossible to fill (particularly at Sao Tomé island in 1985, 1986 and 1987). Some records from the classical tide gauges are of poor quality (at Abidjan-Vridi in 1988, Lomé in 1983 and throughout at St Pedro). No data are available at Pointe-Noire during 1984.

The T/P instruments record the absolute pressure which is the sum of the atmospheric pressure and the water column above the pressure sensor. Seasonal variations of barometric pressure are small in the area (Fig. 1a,b), so data from the T/P sensors are equivalent to those from classical tide gauges after the latter has been corrected for atmospheric pressure variations.

2.4 Processing of the tide gauge data

Each record was plotted, cleaned of obvious errors and submitted to a tidal and spectral analysis. The residuals (observations-predictions) were plotted revealing timing errors in some records. Usually, such errors were obvious and could be attributed to human error, and remedied by shifting the whole (or partial) series by the required number of hours. However, when the residuals increased continuously the dubious part of the record had to be replaced by a prediction.

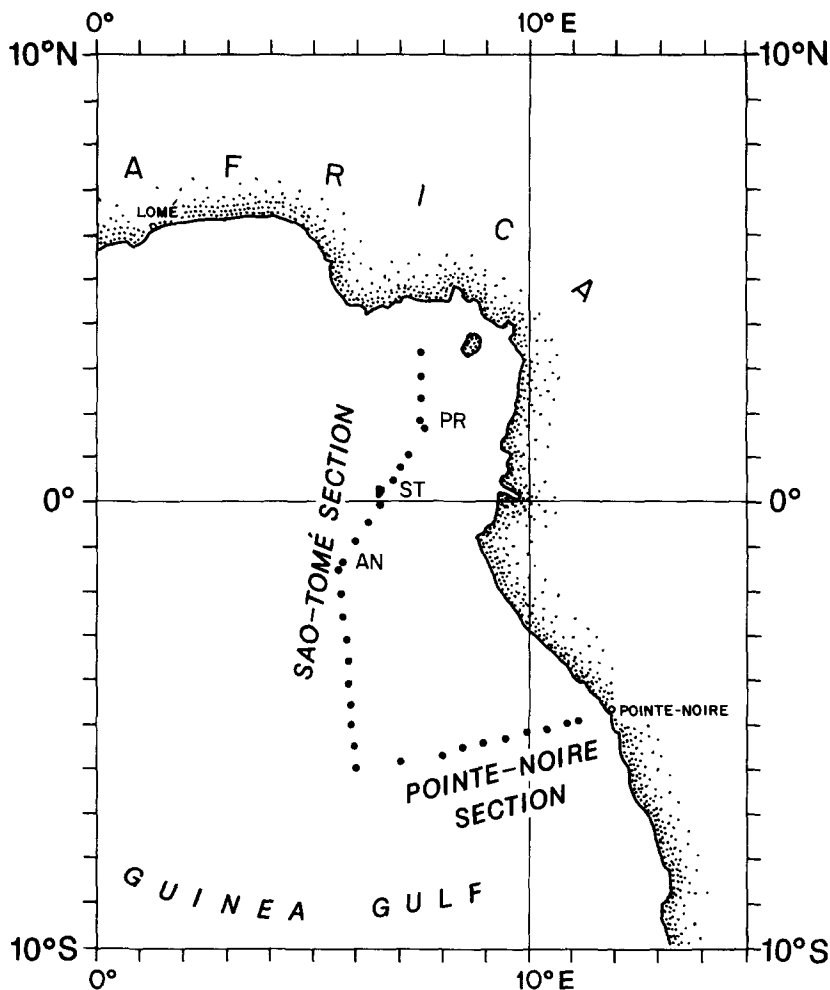


FIG.4. The "TOGA" sections (0-500m), December 1985 - December 1988, Lomé-Pointe-Noire. The 31 stations were worked during 5 campaigns (NIZERY, ORSTOM research vessel). The Sao Tomé section was occupied between 3°30'N and 6°S, the Pointe-Noire section between 6°E and 11°E.

After concatenation of the records n and $n+1$, the junction was checked by plotting the residuals of the concatenated series. Without overlapping data at each recovery-redeployment, step by step checks were performed at each site and then on the total concatenated records. Each resulting hourly record was then filtered over the tidal frequencies (diurnal and semi-diurnal) by a symmetrical filter, which usually spanned 71h of data and had a cut-off frequency of 0.37cpd (2.7 day period) (DEMERLIAC, 1973). Each hourly "detided" series was plotted and the junctions between the different records rechecked. The high-variability in the inertia-gravity (3-6 day periods) to intraseasonal (40-60 day periods) band period was filtered out using a very low-pass filter (half width of 364hr) the gain of which falls from 1 to zero between $f_c=0.00821\text{cd}^{-1}$ (4 months) and $f_c=0.01643\text{cd}^{-1}$ (2 months). The very low-passed series were then 366 hour averaged (15.25 day averages) and plotted to study the seasonal or interannual variability.

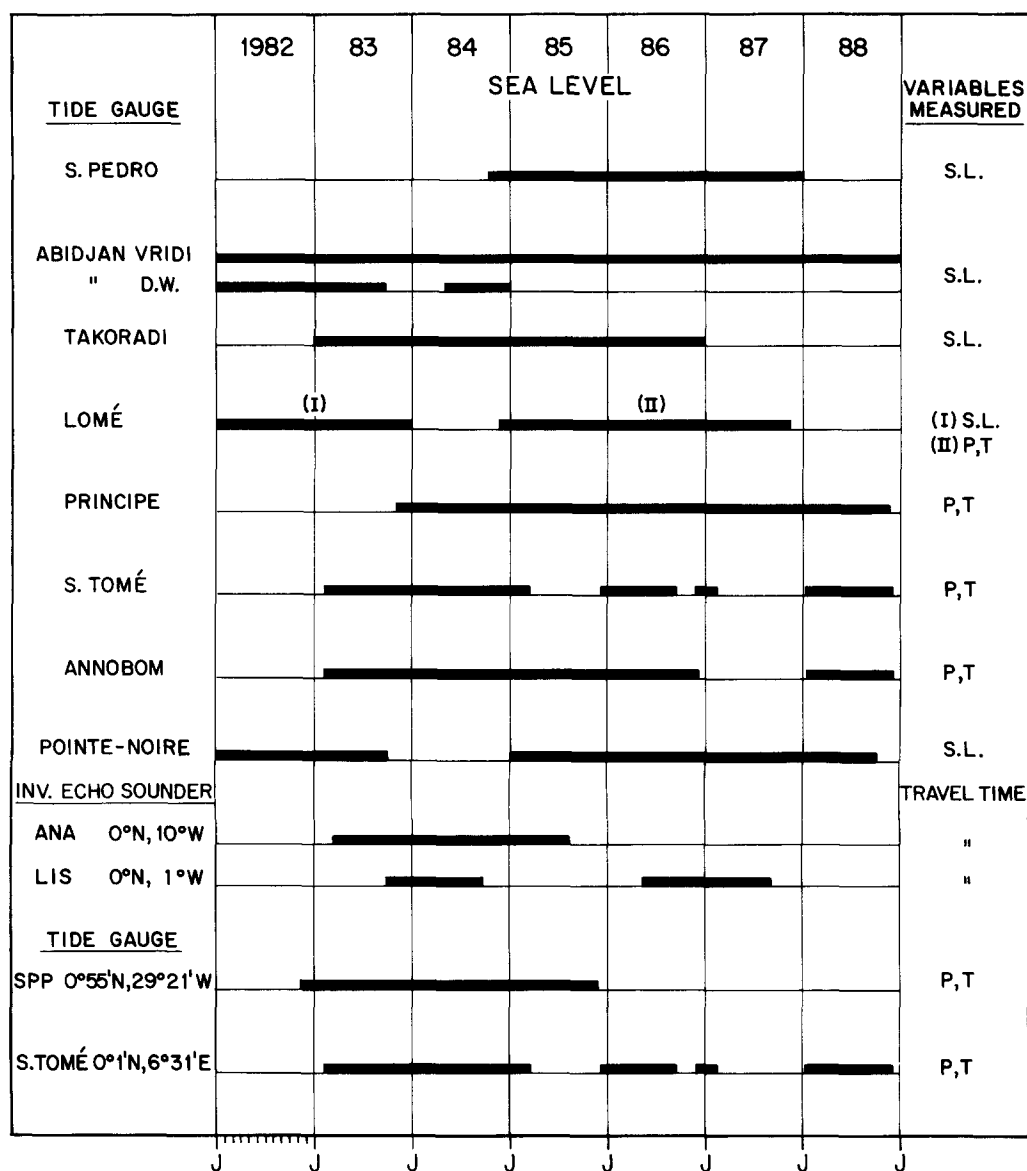


FIG.5. Sea level and Inverted-Echo Sounder (IES) observations in the Gulf of Guinea from 1982 to 1988. (See the positions on Fig.3). Classical tide-gauge: SL; Pressure tide-gauge: P/T (pressure/temperature). Durations are nominal. The exact positions of synoptic sea levels at St Peter and St Paul rocks (SPP) and Sao Tomé and IES (ANA,LIS) observations are also indicated.

3. PROCESSES FOR THE WIND TO FORCE THE OCEAN

3.1 Local forcing by meridional wind at the equator (equatorial Ekman divergence)

The summertime increases in the southerly wind stress in the Gulf of Guinea (the African monsoon) may account for both the cool superficial waters observed from June through early October at and south of the equator, and the strengthening of the eastward Guinea Current observed at 3°-5°N. VOITURIEZ (1981a) hypothesized that south of the equator the southerly winds which increase in April-May at the beginning of the upwelling season (Fig. 2c) induce the oceanic divergence by the process suggested by CROMWELL (1953): the South Equatorial Current (SEC) increases to the south of the equator, but at the equator the flow bends to the North and this results in a divergence of water at about 1°S; upwelling occurs to the east of 10°W, where the wind is mainly southerly. Similarly, at about 1°-2°N there is a convergence of water and a downwelling. Numerical models by PHILANDER (1979), CANE (1979), PHILANDER and PACANOWSKI (1981, 1986a) exhibit the features described by VOITURIEZ, suggesting that the SEC and Guinea Currents are driven by the local meridional winds. However, these increases cannot explain the upwelling observed along the northern coast and the general uplifting of the thermocline throughout the Guinea Gulf (HASTENRATH and LAMB, 1977; MERLE, 1978). Moreover, the meridional absolute velocity at fixed depths (10m and 35m) and the local meridional wind stress observed at the FOCAL equatorial mooring (4°W) were uncorrelated on a monthly time scale. After removal of the fluctuations in the 10- to 20-day period band, the mean flow at 10 and 35m depths over the 19-month record proved to be southwards, counter to the northward wind stress (HOUGHTON, 1989). This result contradicts the hypothesis that the surface local cooling is the result of a locally forced equatorial divergence. Actually, the changes in the depth and thickness of the near surface mixed layer are driven by the depth of the thermocline, not by the local wind. At 4°W in the Gulf the cold SST does not appear until the thermocline shoals, and this occurs well after the local meridional wind has already intensified in April-May. Numerical modelling and observations show that the depth of the thermocline is strongly influenced by the distributed remote wind stress to the west (WEISBERG and TANG, 1987; HOUGHTON, 1989).

3.2 Local forcing by alongshore wind (coastal Ekman divergence)

Along the coasts of the Gulf of Guinea, the local alongshore winds have a weak seasonal variability either along the northern coast or along the southern coast north of 10°S (Fig. 2b). However, a strong seasonal upwelling occurs each year along these coasts between June and October. SST cools by about 5°C between 4°E and 8°W along the northern coast and between 1°S and 10°S along the southern coast. This surface cooling is not limited to the thin surface mixed layer and is observed as deep as 500m; for example, in 1971 the dynamic heights (0/90dbar) dropped by 10-15dyn-cm over the Ivoirian shelf between Cape Palmas (8°W) and Cape Three Points (2°W) between June and July (Fig. 6).

Calculations of Ekman divergence (VERSTRAETE, 1970a) give vertical velocities (0.7mday⁻¹) which are much smaller than those inferred from the observed vertical displacement of the thermocline and isotherms. PICAUT (1983) computed an upward phase propagation of the upwelling event of 7mday⁻¹ from 350m to 50m from the isotherm vertical displacements at a station situated 40km off Abidjan. HOUGHTON (1976), BAKUN (1978) and PICAUT (1983) all failed to find a correlation between the near shore SST variations and local winds. Although CLARKE (1978) suggested a local generation of the seasonal coastal upwelling, the coupling

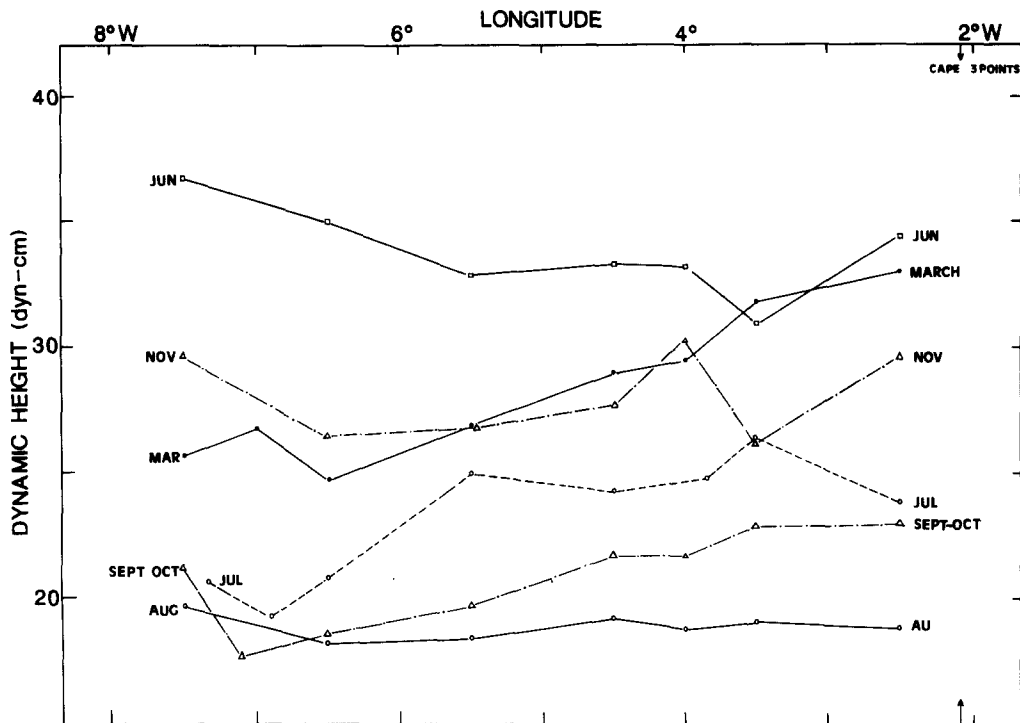


FIG.6. Dynamic heights (0/90dbar) along the edge (200m isobath) of the Ivoirian shelf from March to November 1971.

coefficient (which measures how efficiently wind stress couples to the ocean) of his reduced-gravity model is much too big to be realistic. PHILANDER (1979) and PHILANDER and PACANOWSKI (1981) showed that southerly winds could cause low SST along the southern coast, but their assumed wind field and coastal cooling were unrealistic. Cross equatorial winds do contribute to the intensification of the Guinea Current during the summer, but the upwelling they generate along the northern coast is insignificant.

In summary, all these previous studies show that neither at the equator nor along the coasts can Ekman divergence explain a significant part of the observed annual upwelling in the Gulf of Guinea.

3.3 Remote forcing generated by zonal wind changes along the equator

During the FINE workshop (FGGE/INDEX/NORPAX Equatorial workshop, June 27 - August 12, 1977) at UCSD, La Jolla, California, the French participants presented new long time series of SST, sea level, hydrography and Meteorological data collected at 27 different coastal stations along the coast of the Gulf of Guinea (review by VERSTRAETE *et al*, 1980). Based on the theoretical works of MOORE (1968) and LIGHTHILL (1969), MOORE, HISARD, McCREARY, MERLE, O'BRIEN, PICAUT, VERSTRAETE and WUNSCH (1978) were the first to suggest that throughout the Gulf of Guinea a significant fraction of the annual upwelling was forced remotely by a sudden increase of the easterly wind in the central-western equatorial Atlantic. This remote forcing

generates an equatorially trapped Kelvin wave which travels into the Gulf and is reflected, first polewards along the coast of Africa as coastal Kelvin waves, and then back westwards as Rossby waves. Subsequently, an extremely simple linear model of the equatorial Atlantic was formulated by O'BRIEN, ADAMEC and MOORE (1978) and ADAMEC and O'BRIEN (1978) which showed the physical mechanism in a theoretical context. The critical parameter used in their two-layer model was the equivalent depth. A value of 10cm was chosen as being appropriate for the coastal region, and the trapping scales used were 225km at the equator (corresponding to the second baroclinic mode, Table 2) and 100km along the northern coast. In both studies, a westward impulse of $0.25 \text{ dyne cm}^{-2}$ was applied in the western equatorial Atlantic. At days 55-70, the progression of the upwelling Kelvin wave along the equator and along the coasts raised the thermocline by 45m and lowered sea levels by nearly 10cm, values consistent with observations.

The authors recognised that their model and its forcing are undoubtedly too simple. It was analogous to the explanation for the El Niño event in the eastern Pacific proposed by McCREARY (1976) and HULBERT, KINDLE and O'BRIEN (1976). These early theoretical and numerical works were very important in demonstrating the power of the "remote forcing" concept in explaining phenomena observed in the eastern equatorial oceans, such as El Niño events in the Pacific and the seasonal upwelling in the eastern Atlantic. However, on the basis of a single baroclinic mode, the significance of impulse forcing remained questionable (CANE and SARACHIK, 1979; PHILANDER, 1981a,b).

The continuous record of *in situ* winds during 1979-1988 at St Peter and St Paul rocks (RSPP) show that each year the intensification of the westward wind stress τ_x is very rapid. For example, KATZ (1987a) identified strong pulses on April 11 1983 and on May 21 1984. Although these pulses do not appear in the climatology because of the averaging process, there is a strong seasonal signal in τ_x in April-May in the area $10^\circ - 30^\circ\text{W}$ (Fig. 2a). Thus the seasonal increase of τ_x can be considered as an initial forcing factor. PHILANDER and PACANOWSKI (1981) used seasonal impulsive forcings and their simulations were reasonably realistic (KATZ and GARZOLI, 1982).

The main problems with reduced-gravity models are their inability to generate vertical phase propagation (McCREARY, 1984) and the dynamics of baroclinic modes excitation is poorly understood (CANE and SARACHIK, 1979). A very important feature of the upwelling signal is a vertical phase propagation observed both in the eastern equatorial Pacific (LUKAS, 1981) and in the Gulf of Guinea (PICAUT, 1983; VERSTRAETE, 1985b), and only continuously stratified models can simulate this feature (McCREARY, 1981, 1984).

3.4 Equatorial wave theory

The state of knowledge of the time-dependent dynamics of the ocean at low latitudes in response to large-scale winds is based on the theory developed by MOORE (1968) and LIDTHILL (1969). Reviews given by MOORE and PHILANDER (1977), GILL (1982), McCREARY (1981, 1984) and PHILANDER (1990) provide the background for studies of the dynamics of equatorial waves, so here only a brief summary of the mathematical theory is necessary.

The equations governing equatorial long waves were first developed by LAPLACE for shallow waters on a rotating sphere. Their solution consists of an infinite set of meridional Hermite modes L propagating horizontally (for each vertical standing baroclinic mode n) or an infinite set of vertically propagating waves (each wave being associated with a meridionally trapped Hermite mode L). Waves associated with a considerable number of vertical modes superimpose in such a way that energy and phase propagate both vertically and horizontally (McCREARY, 1984). On the dispersion diagram for equatorially trapped modes, there are four classes of waves (frequency σ , wavenumber k):

- The Kelvin wave ($L=-1$ and $k \geq 0$), non dispersive
- The high-frequency Inertia-Gravity waves ($L \geq 1$ and $k \geq 0$ or $k \leq 0$)
- The low-frequency Rossby waves ($L \geq 1$ and $k \leq 0$)
- The mixed Rossby-gravity (or Yanai) waves ($L=0$ and $k \geq 0$ or $k \leq 0$)

Phase propagates eastwards for the Kelvin wave and westwards for Rossby modes, while Inertia-Gravity (I/G) and Rossby-gravity waves may propagate in either direction. Energy propagates eastward for Kelvin and Rossby-gravity waves, but in either sense for I/G and Rossby waves. The equatorial scalings are $(c_n/\beta)^{1/2}$ for the zonal wave number k , and $(\beta c_n)^{1/2}$ for the frequency σ .

The horizontal (L) and vertical structures (n) are coupled via a “separation constant” $c_n = (gh_n)^{1/2}$ where h_n is the “equivalent depth” of a stratified ocean. The trapping latitudes $[c_n(2L+1)/\beta]^{1/2}$ depend on the vertical mode n and the horizontal mode L ($L > 1$). For negative wave numbers, the higher the meridional mode, the further poleward the Rossby wave extends. At 6°E , the trapping latitudes for the first Rossby meridional mode ($L=1$) are 553km (first baroclinic mode) and 398km (second baroclinic mode).

In the frequency band:

$$\left[\frac{L+1}{2} - \frac{L}{2} \right] (\beta c_n)^{1/2} \leq \sigma \leq \left[\frac{L+1}{2} + \frac{L}{2} \right] (\beta c_n)^{1/2}$$

the wave numbers k are complex and the imaginary part:

$$k(\text{Im}) = \pm(2L+1-\sigma^2-1/4\sigma^2)^{1/2}(\beta/c_n)^{1/2}$$

correspond to exponential decay of a coastally trapped motion along a continental boundary (sign $+$ ($-$) is for exponential decay toward the west (east), respectively). In this frequency range neither I/G nor Rossby waves can exist. For the first (second) baroclinic mode at 6°E (Tables 1,2), the separation constant is 233cms^{-1} (121cms^{-1}) and the reflectionless gap extends from 5.8 to 34.0 day-periods (8.1 to 47.2 day-period). Hence, wave energy in the 1 to 5 week period range (1 to 7 week period range) can only propagate eastwards in the form of Kelvin wave and Rossby-gravity waves. When it impinges on a continent, the energy cannot be reflected because no I/G or long Rossby waves are available in the reflectionless gap. Therefore, the energy must propagate polewards away from the equator along the coast as trapped waves (MOORE, 1968; CLARKE, 1983). Because the offshore scale of the trapped waves is larger than the scale of the coastal topography, they behave like internal Kelvin waves trapped by a vertical wall (ALLEN and ROMEA, 1980).

The meridional structure for sea level, temperature and pressure has even (odd) symmetry about the equator depending upon whether the meridional mode number L is odd (even). For equatorial Kelvin wave, the sea surface displacement is maximum at the equator and falls off symmetrically to the north and south with a Gaussian shape. For the typical stratifications found in the equatorial Atlantic at 35°W and 6°E during 1982-84, the speed of Kelvin wave decreases from 241 (143) to 233 (121) cms^{-1} for the first (second) baroclinic mode (Tables 1 and 2).

If low vertical standing modes exist (rather than vertically propagating waves) their baroclinic structure impose a signature on sea level variations. The vertical structure functions for the first three horizontal velocity, sea level and temperature baroclinic modes during 1982-84 at 35°W and 6°E show that the first mode crosses zero at 1300m both at 35°W and 6°E (Fig.7). In the upper 500m, the zero crossings of the second mode deepens from 130m at 35°W to 290m at 6°E . The vertical structure of the second baroclinic mode suggests that temperatures below 130m (at 35°W)

TABLE 1. Parameters associated with baroclinic vertically standing modes for the mean density profile at 0°, 35°W (1982-1984); h: equivalent depth; c: phase speed; L: internal Rossby radius; T: time scale at the equator

mode	h(cm)	c(cms ⁻¹)	L(km)	T(days)
1	59.4	241	326	1.56
2	20.8	143	251	2.03
3	8.2	90	199	2.56
4	4.2	64	168	3.03
5	2.4	48	145	3.50
6	1.8	42	136	3.75
7	1.2	35	124	4.10
8	1.0	31	117	4.35
9	0.7	27	108	4.71
10	0.6	24	103	4.92

$$c_g = (gh_g)^{1/2}$$

$$L = (c_g/\beta)^{1/2}$$

$$T = (\beta c_g)^{-1/2}$$

TABLE 2. Parameters associated with baroclinic vertically standing modes for the mean density profile at 0°, 6°E (1982-1984); h: equivalent depth; c: phase speed; L: internal Rossby radius; T: time scale at the equator

mode	h(cm)	c(cms ⁻¹)	L(km)	T(days)
1	55.4	233	320	1.59
2	14.9	121	231	2.21
3	6.6	81	188	2.71
4	4.2	64	168	3.03
5	2.8	53	152	3.35
6	1.8	43	137	3.72
7	1.4	37	127	4.01
8	0.9	30	115	4.41
9	0.8	28	111	4.58
10	0.6	25	105	4.87

and 290m (at 6°E) and sea level will have the opposite phases; conversely, near sea surface temperature and sea level will be out of phase because the strong deep baroclinic signal is first recorded by the sea level when the sea surface temperature variations are mainly the result of the vertical displacements of the shallow thermocline.

Because the equatorial waves travel with high group speeds, they are of paramount importance in the oceanic adjustment to a change in the winds. The fastest waves are the Kelvin waves which carry the energy eastwards at c_g . The long Rossby waves are nearly non-dispersive and propagate the energy westwards at the group (or phase) speed $c_g/(2L+1)$. In contrast, the Inertia-Gravity (I/G), Rossby-Gravity and short Rossby waves are relatively unimportant; these can be filtered out by making the "long-wave" approximation (PHILANDER, 1981a,b, 1990). The adjustment time of the equatorial ocean is the sum of the time a Kelvin wave takes to propagate eastwards plus the time a long Rossby wave takes to propagate back westwards across the basin. This is estimated to be 150 days and 450 days in the Atlantic and the Pacific respectively (CANE, 1979; PHILANDER, 1981a).

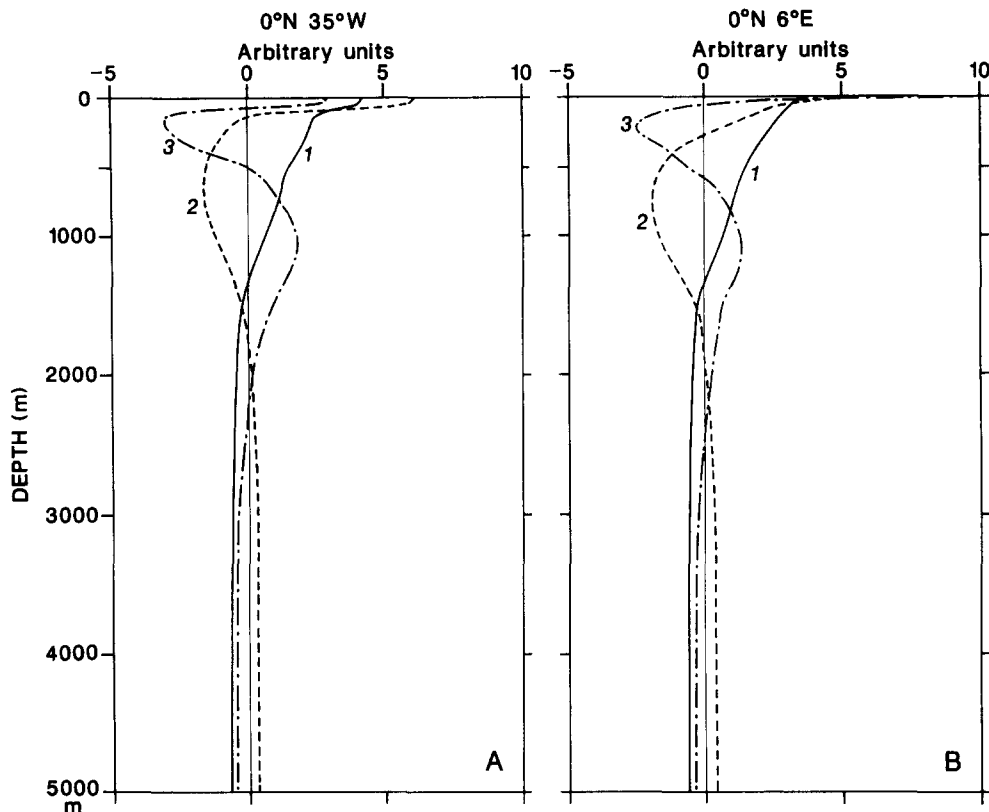


FIG.7. The vertical structure functions for the first three horizontal velocity and pressure (sea level) baroclinic modes at (a) 0° , 35°W and (b) 0° , 6°E ; computed from the FOCAL density mean profile (0-500m) and the GEOSECS data (500-5000m).

We are concerned here with the existence of low vertical modes as the cause for sea level variations. In the Pacific, WUNSCH and GILL (1976) and ERIKSEN (1982) have shown that it is possible for I/G waves to have standing vertical modes in areas where the bottom is flat. However, this is not the case in the Atlantic. GARZOLI and KATZ (1982) and GARZOLI (1987) give evidence of forced I/G waves at 28°W ; WEISBERG, HORGAN and COLIN (1979) have observed equatorially trapped Rossby-gravity waves at 4°W which had a downward energy flux. According to PHILANDER (1981a), the adjustment of the upper equatorial ocean to zonal wind stress impulses is effected by vertical modes trapped in and above the sharp shallow thermocline by internal reflection. The second baroclinic mode has a large amplitude primarily above the thermocline (Fig.7) and its vertical structure is similar to that of the intense equatorial currents in the upper ocean. So this mode is likely to be excited preferentially when those currents change in response to a change in the winds. The value of the equivalent depth for this mode (21cm at 35°W , 15cm at 6°E , Tables 1 and 2) corresponds to the equivalent depth usually chosen in shallow-water models in which the thermocline is the interface. This reinforces our expectation that it is the second baroclinic mode which is the most important in the adjustment of the upper equatorial ocean to basin-wide changes in the wind (PHILANDER, 1990).

4. HYDROGRAPHY OF THE EQUATORIAL ATLANTIC

Following the GATE programme in 1974 and GEOSECS Atlantic expeditions in 1972-73, a description of the tropical Atlantic water masses (0-3000m, 10°S - 30°N), excluding the eastern Gulf of Guinea, was given by EMERY and DEWAR (1982). It is unnecessary to list all the studies by the ORSTOM group in the Gulf of Guinea, apart from the descriptions of the seasonal variability given in MERLE's (1978) atlas, and MERLE and ARNAULT (1985), and also the papers on the hydrography, circulation, nutrients, dissolved gases in the Gulf included in a special issue of "Océanographie Tropicale" (1983, Vol.18). The water masses and the hydrographic features of the equatorial undercurrent and the equatorial thermostad are now examined in relation to the equatorial and coastal upwelling.

4.1 Water masses and essential hydrographic features

Below the surface waters (temperature above 26°-28°C), the mean temperature-salinity (T/S) diagrams of the upper 500m of the tropical Atlantic show there to be two water masses (Fig.8): The North and South Atlantic Central Water (NACW and SACW). Off Natal (5°-7°S, 34°-36°W), St Peter and St Paul rocks (RSP), along the equator at 10°W, 4°W, 6°E and off Pointe-Noire (5°S), SACW can easily be defined between 100 and 500m. For the sake of comparison, NACW off Martinique is presented in Fig.8. A detailed description of the mean T/S relationship and its seasonal variation off Abidjan and Pointe-Noire provided a new insight into the distribution of upper and mid-depth water masses in these areas (VERSTRAETE, 1985b, 1987). Off Abidjan, 217 Nansen casts were done from 1957 to 1964 by ORSTOM, of which 80 reached a depth of 1400m. Four water-types were defined (salinity in psu, Fig.9): (1) the Tropical surface water (28°C, 35.00); (2) SACW (18°C, 35.60); (3) the Antarctic Intermediate Water (AAIW) (5°C, 34.55); (4) the North Atlantic Deep Water (NADW) (3.5°C, 35.00) based on characteristics given by SVERDRUP, JOHNSON and FLEMMING (1942) and EMERY and DEWAR (1982).

According to COCHRANE, KELLEY and OLLING (1979), three subthermocline flows carry the SACW eastwards between 5°S and 5°N, branching off the North Brazil current. At 4°W, the EUC and the South Equatorial Countercurrent (SECC) are permanent features (HISARD, CITEAU and MORLIERE, 1976; MOLINARI (1982); MOLINARI, VOITURIEZ and DUNCAN (1981); VOITURIEZ, 1983a,b). Following TSUCHIYA (1975), MOLINARI (1982) distinguished the South Equatorial Undercurrent (SEUC) at 3°-5°S and the SECC at 7°-9°S. The North Equatorial Undercurrent does not penetrate into the Gulf where only the shallow Guinea current is observed between 2° - 3°N and the African coast. During 1982-84, the EUC and the SEUC extended across the Atlantic to at least 6°E.

The most striking of these eastward subsurface flows is the EUC which has been studied in great detail during the last 30 years, and the recent FOCAL/SEQUAL programmes (1982-84) have added significantly to observations on this major geophysical flow. Three hydrographic features related to the EUC appear consistently in each section (Figs 10,11): (1) a high-salinity core (Figs 10 and 11b), found above the 20°C isotherm in the area 35°-10°W and below the 20°C isotherm east of 4°W (Fig.11a); (2) a strong vertical spreading of the isotherms from the core of the undercurrent (Figs 10a,c,e,g); and (3) a large thermostad of SACW between 12° and 16°C (Figs 10, 11), this water mass (Figs 10a,c,e,g) being clearly identified on T/S diagrams (Figs 8 and 9).

During 1982-84, the isolated high-salinity core appears clearly in each of the mean vertical sections from 35°W to 6°E (Figs 10,b,d,f,h). It was found above the zonal velocity maximum from 35°W to 4°W and at the same level at 1°E and 6°E. At 35°W, 4°W and 1°E, the high-salinity core

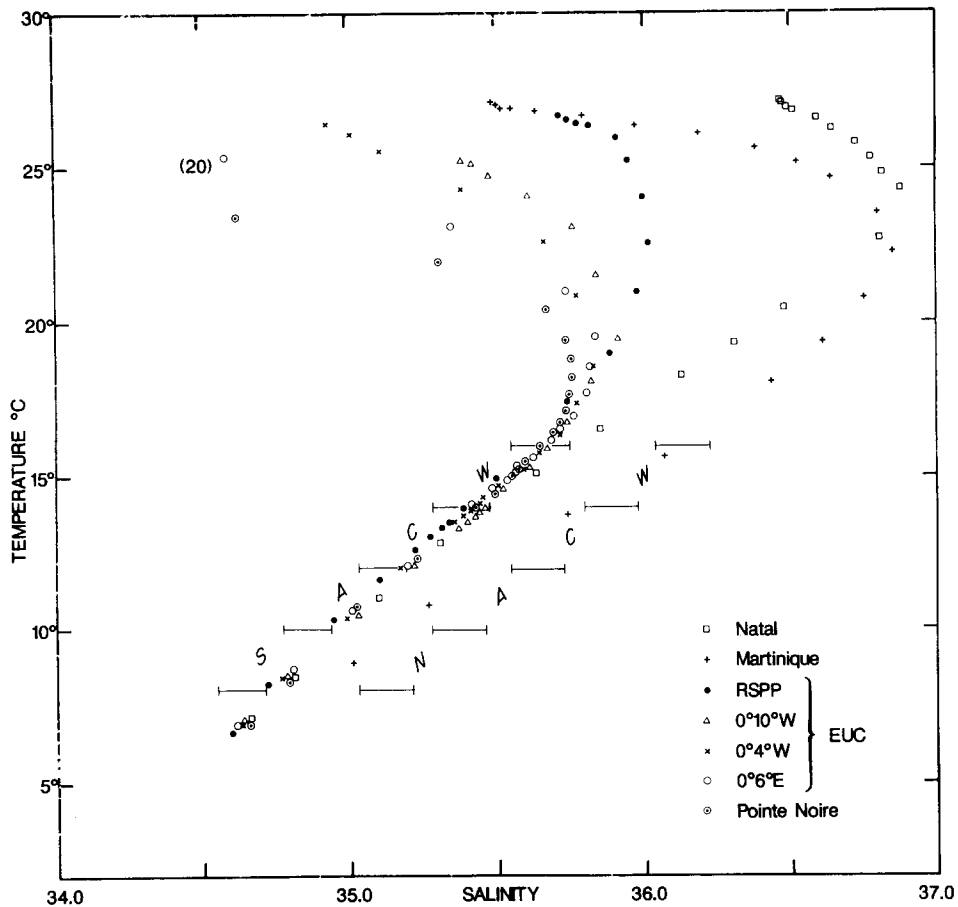


FIG.8. Thermocline water masses in the tropical Atlantic ocean: central waters are those of the North Atlantic (NACW) and those of the South Atlantic (SACW), between 8°C and 16°C. At the same temperature, the southern waters are less saline than the northern ones. The horizontal bars show the ranges of salinities at given temperatures after SVERDRUP *et al* (1942).

was found to the South of the equator at 0°30'S, shoaling from 80-90m at 35°W to 40-50m at 1°E. At 6°E, the high-salinity core straddled the equator between 1°S and 1°N, and a second isolated core was observed at 3°S at 40m. A tongue of high-salinity water protrudes northwards at the depth of the thermocline in each section, which is separated from the core of the EUC by water of low salinity. The mean salinity in the core of the EUC decreased from about 36.3 at 35°W to about 35.9 at 6°E.

The mean sections at 6°E and 1°E show a region of relatively high salinity at about 60 to 100m, from 2°N to 3°N (Figs 10f, 10h). The salinity in this tongue seems to decrease from 6°E to 1°E, but the tongue is not apparent at 4°W, probably averaged out. However, this salinity maximum appeared clearly at 5°E in May 1973 (MORLIERE, HISARD and CITEAU, 1974) and in May 1983, August 1983, May 1984 and July 1984 at 6°E, 1°E and 4°W during FOCAL (HÉNIN, HISARD and PITON, 1986). These results suggest that there is continuity between the EUC and the Guinea Undercurrent between 2°N and 4°N.

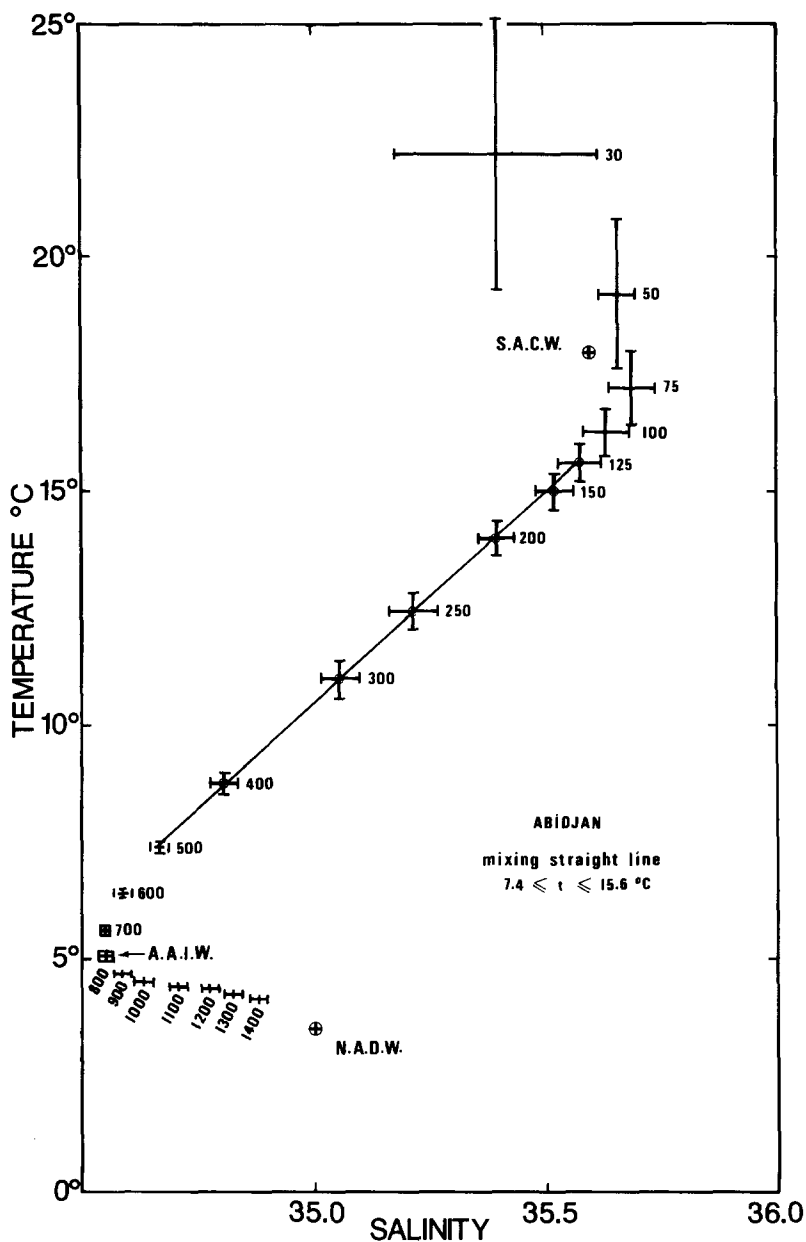


FIG.9. Mean T-S diagram off Abidjan between 30m and 1400m. Each dot gives the mean temperature and salinity of the 12 monthly means. Figures indicate the depth in metres. Standard deviations are shown at each level for temperature and salinity. The mixing straight line is between 7.4°C (500m) and 15.6°C (125m). SACW: South Atlantic Central Water; AAIW: Antarctic Intermediate Water; NADW: North Atlantic Deep Water.

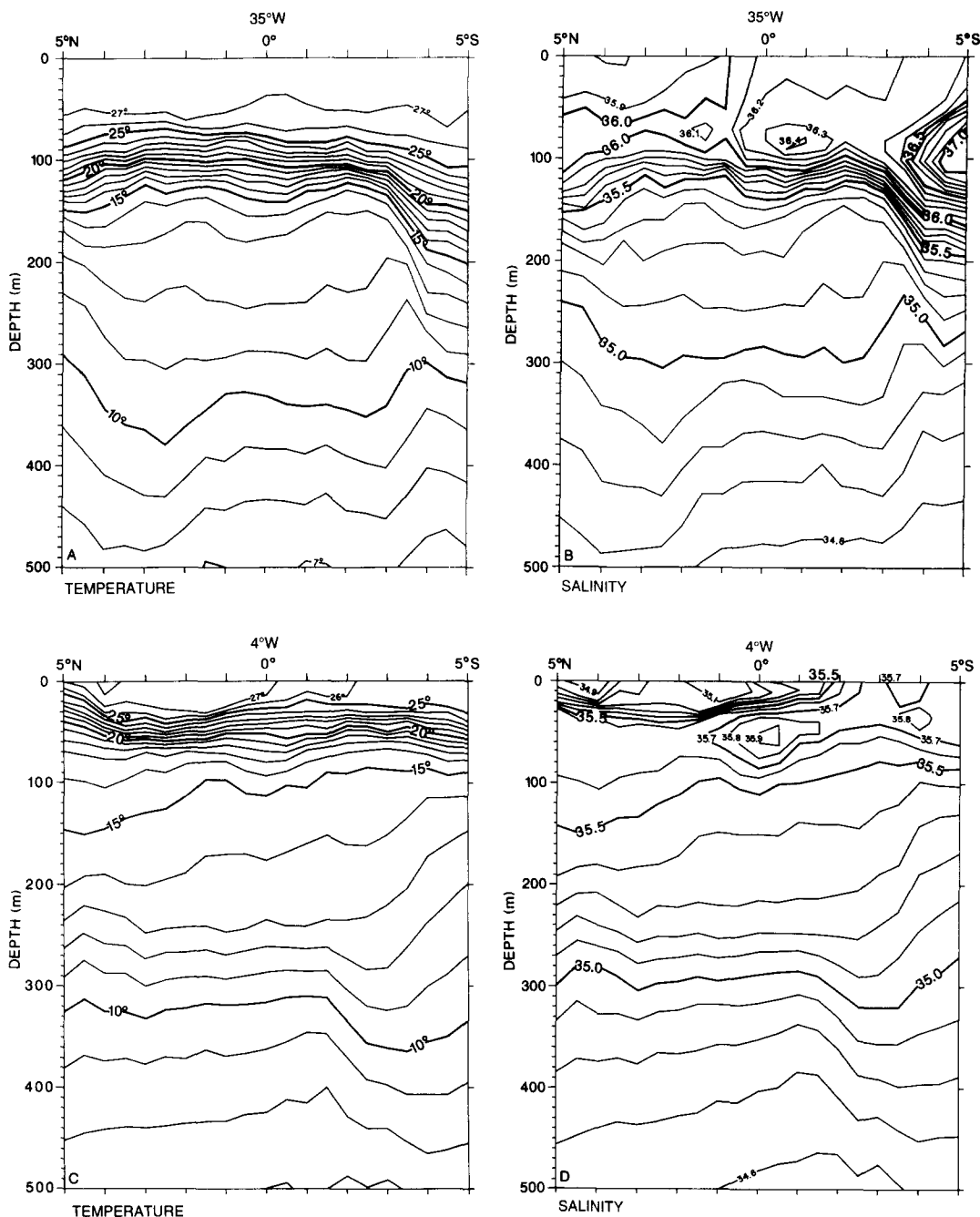


FIG.10. (above and overleaf) Sections (0-500m) of the mean temperature and salinity at (a,b) 35°W, (c,d) 4°W, (e,f) 1°E and (g,h) 6°E computed from the FOCAL data (1982-1984). The latitudinal grid of the stations is 1/2°, between 5°N and 5°S. Note the vertical spreading of the isotherms at the equator in the upper 150m due to the equatorial under current.

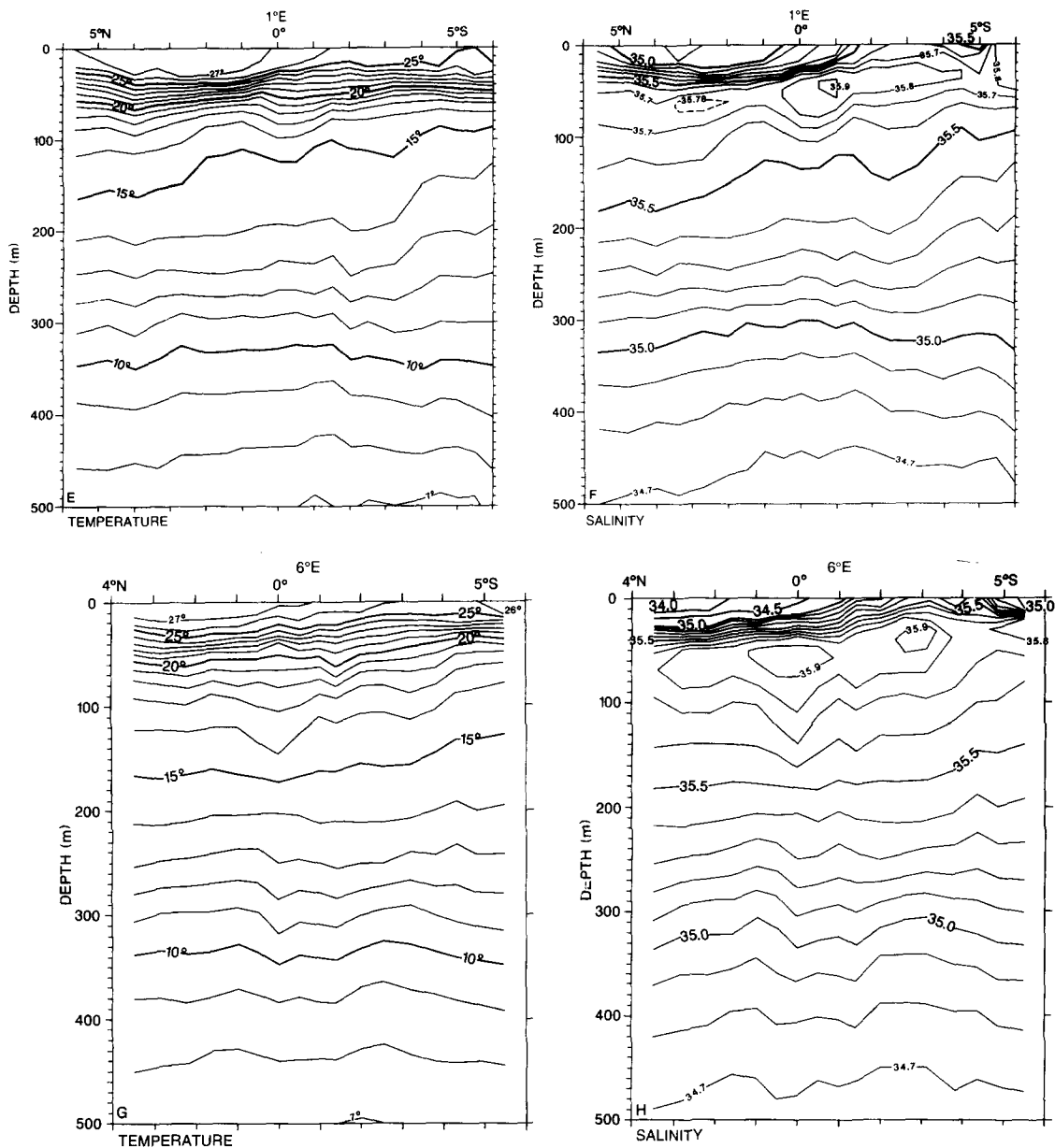
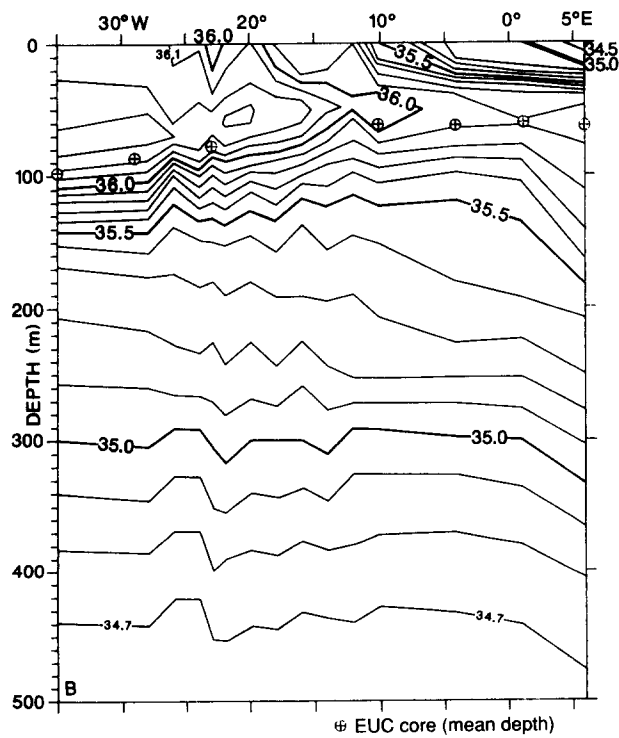
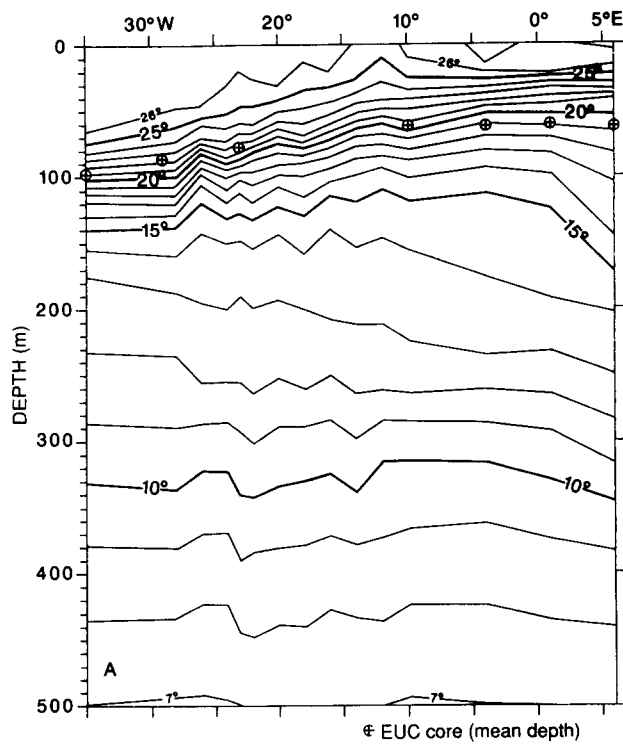


FIG.10 (continued).

FIG.11a. (opposite) Section (0-500m) of the mean temperature along the equator between 35°W and 6°E in 1982-1984. To the west (east) of 10°W the core of the EUC maximum zonal speed (near the 20°C isotherm) lies in the upper (lower) part of the thermocline.

FIG.11b. (opposite) Same as Fig.11a for salinity. To the west (east) of 10°W, the core of the EUC maximum zonal speed lies below (inside) the high-salinity core.



From direct PCM measurements of current made during the eight FOCAL cruises, we have computed the means of the zonal speed, the depth and the volume transport of the EUC at 35°W, 29°W, 23°W, 10°W, 4°W, 1°W and 6°E. From 35°W to 6°E, the average of the maximum speed in the EUC core decreased from about 80 cm s⁻¹ to 30 cm s⁻¹; the depth of the core shoaled from 100 m to 60 m and its volume transport (inside the 10 cm s⁻¹ isotach) decreased from about 16 Sverdrup (1 Sverdrup = 10⁶ m³ s⁻¹) to 2.7 Sverdrups. The standard deviations are 7 Sverdrups at 35°W (6 samples) and 2 Sverdrups at 6°E (8 samples). It is particularly interesting to examine the mean sections of the observed zonal velocity at 4°W and 6°E (Figs 12a,b). At 4°W, the EUC is strong, greater than 50 cm s⁻¹ at 60 m, and nearly centered at the equator. At 6°E, the EUC is very weak at the equator (+7 cm s⁻¹ at 80 m) because it is split into two separate cores: one at 1°N of +25 cm s⁻¹ at 60 m, and another one at 0°30'S, weaker and deeper (+15 cm s⁻¹ at 100 m). The island of Sao Tomé blocks the EUC at the equator and splits it into these two isolated cores. At 0°30'N, the northern core (40-100 m) carried a salinity maximum of 36.0 at 60 m embedded within a tongue of high salinity (35.9) water extending between 1°N and 1°S. At 3°S, 6°E there is a separate high salinity core (35.9) corresponding to the SEUC (+30 cm s⁻¹ at 20 m). This subsurface flow also appears at 4°W between 3° and 5°S, at 20 m to 300 m (Fig. 12a).

On the 4°W zonal velocity section (mean transport 9 Sverdrups, standard deviation 3.9 Sverdrups for 8 samples), the EUC appears nearly completely surrounded by westward flows; the two westward subthermocline flows at 2° - 3°S and 1°30'N - 4°N are important and correspond to the two large thermostads flanking the EUC (Fig. 10c and 12a).

4.2 The seasonal cycle along the equator

At the equator, a strong annual signal appears nearly simultaneously during March-April-May in the high-salinity core at 10°W, 4°W and 6°E (Figs 13 and 14). In July-August this salinity maximum disappears, but reappears during October-January. At 10°W, 4°W and 6°E vertical profiles of temperature and salinity show a strong cooling from April to August between 0 and 200 m and a sharp decrease of the salinity maximum, suggesting a strong upwelling (Figs 13 and 14). Two hydrographic campaigns east of 6°E in May and December 1986 provided four typical salinity sections (Figs 15a,b). In May 1986 (upwelling onset), at 6°E a large salinity maximum (above 36.1) occurred between 1°N and 4°30'S whereas in December 1986 (warm season), the equatorial salinity maximum was reduced (35.8) but was more extended off Pointe-Noire than in May 1986. This is probably a consequence of the absence of both upwelling and the poleward undercurrent along the coast during the warm season (WACONGNE, 1988).

The seasonal variation of the thermostad at 0°N, 6°E (between the 14° and 16°C isotherms) appears clearly on Fig. 14a. The vertical salinity gradient is weaker in the thermostad than above and below it. Thus, the thermostad is also a halostad. The thermostad salinity ranges from 35.4 to 35.7 (Fig. 14b). From January to April the thermostad thins as a result of the depression of all the isotherms above 14°C. This is the annual downwelling which is observed as well off Pointe-Noire and Abidjan, and is as important an event as the upwelling in the seasonal cycle of the eastern equatorial Atlantic. Beginning in May, the thickness of the thermostad increases until July-August followed by a decrease to the end of the year (Fig. 14). The increase during the May-August upwelling period is a result of the uplifting of the isotherms above 15°C and not to a change in the depth of the base of the thermostad (the 14°C isotherm). Clearly, the increase in thickness of the thermostad in April-May at 6°E is directly related to the annual upwelling event. The spreading of the isotherms at the bottom of the thermocline, concomitant with an increase in the salinity maximum in February-May (Fig. 14a,b) suggests that there is an increase of the EUC transport at 6°E as well as at 10°W and 4°W (Fig. 13). The spreading of the thermostad in April-May,

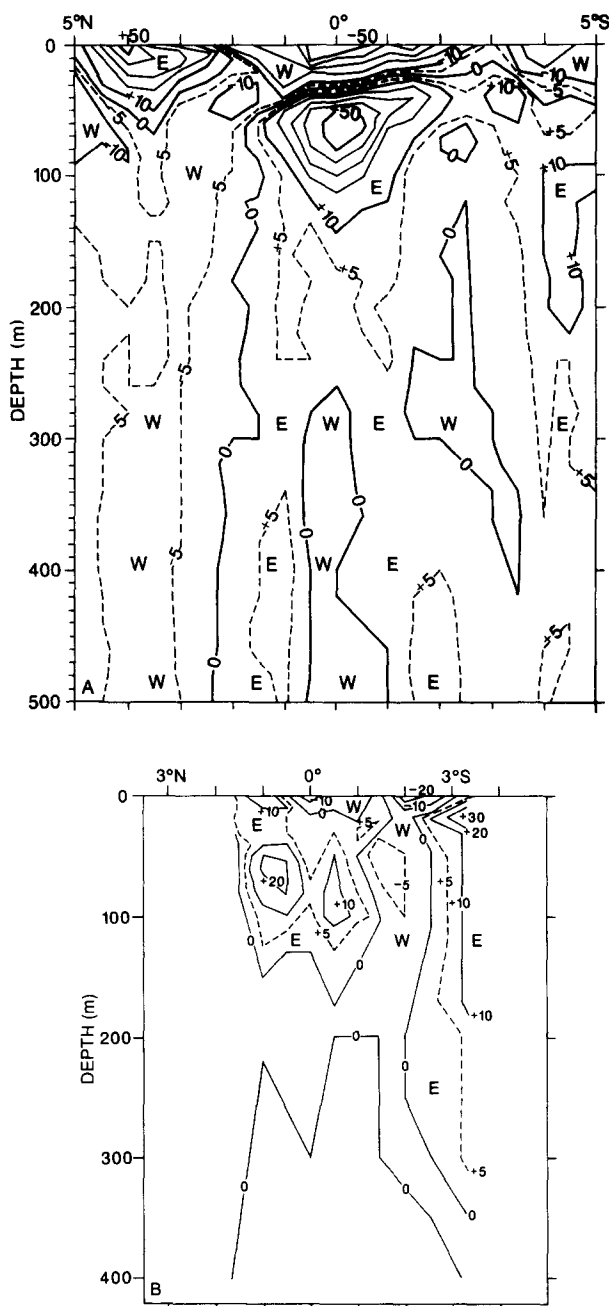


FIG.12. Sections of the mean zonal speed from eight direct observations of current by means of a profiler during 1982-1984. (a) between 5°N and 5°S at 4°W (0-500m). The core of the EUC lies at 60m. The South Equatorial Under Current (SEUC) flows at 3° - 5°S (50-300m). Note the two westward subthermocline flows at 2° - 3°S and 1°30' - 3°N. (b) between 1°30'N and 3°S at 6°E (0-400m). Two isolated EUC cores appear: at 1°N, depth 60m; and at 0°30'S, depth 100m. The SEUC reaches the surface at 3°S. Eastward currents (E) are positive; westward currents (W) negative.

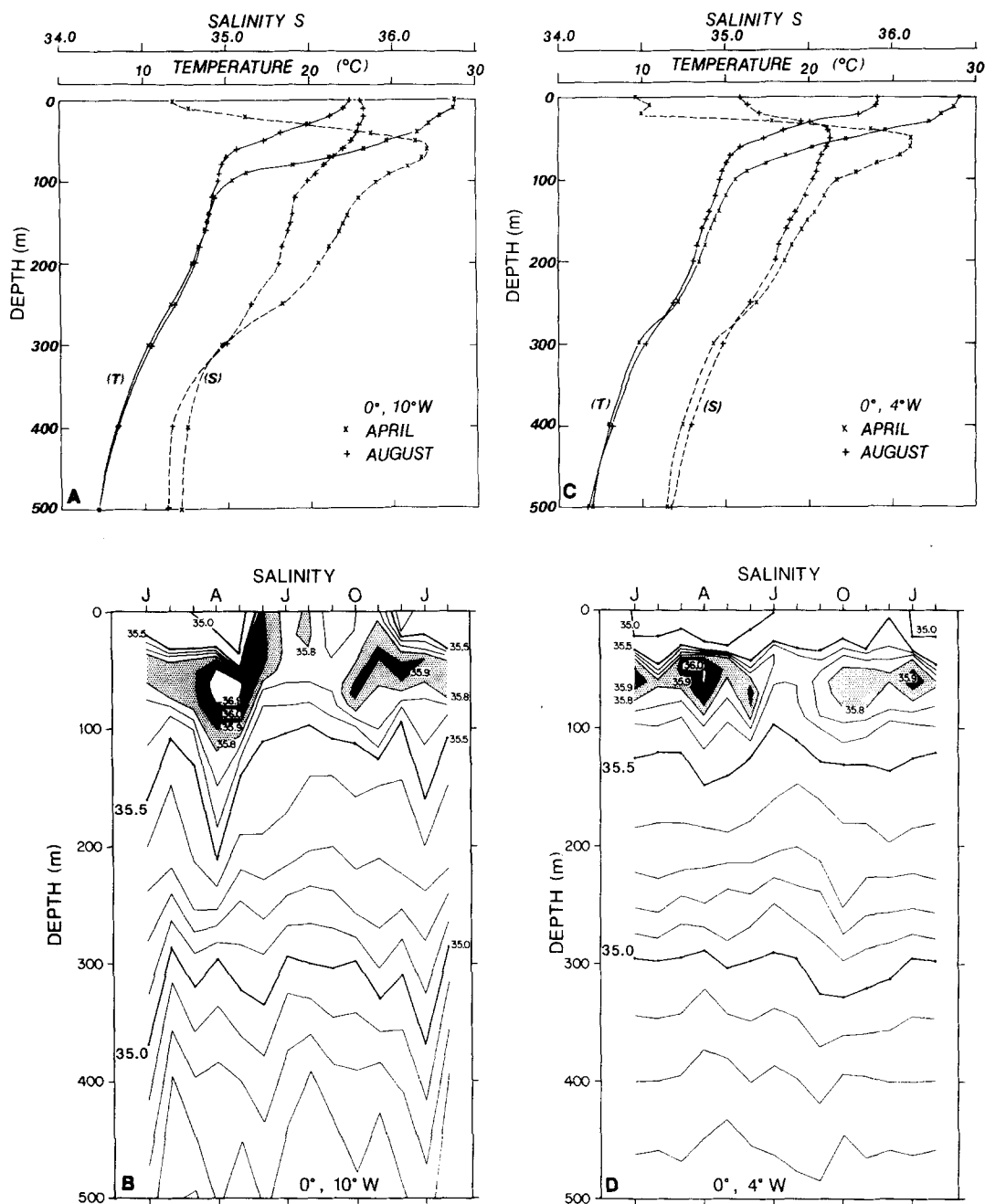


FIG.13. Temperature and salinity-depth profiles (0-500m) at the equator at (a) 10°W and (c) 4°W, during the warm/cold seasons (April/August). Annual cycles in Salinity (0-500m) along the equator at (b) 10°W and (d) 4°W. Note the strong signal between March-April and July-August in the upper 150m.

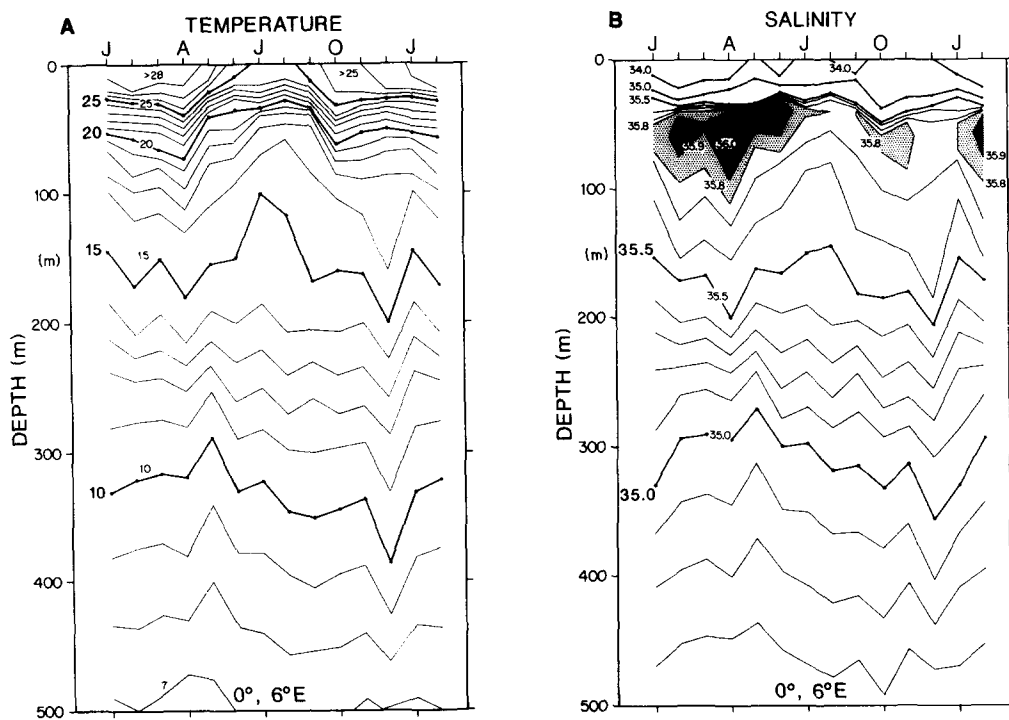


FIG. 14. Annual cycles of (a) temperature and (b) salinity at 0°, 6°E in the upper 500m. Note the strong cooling in April-May concomitant with lower salinity waters below 250m.

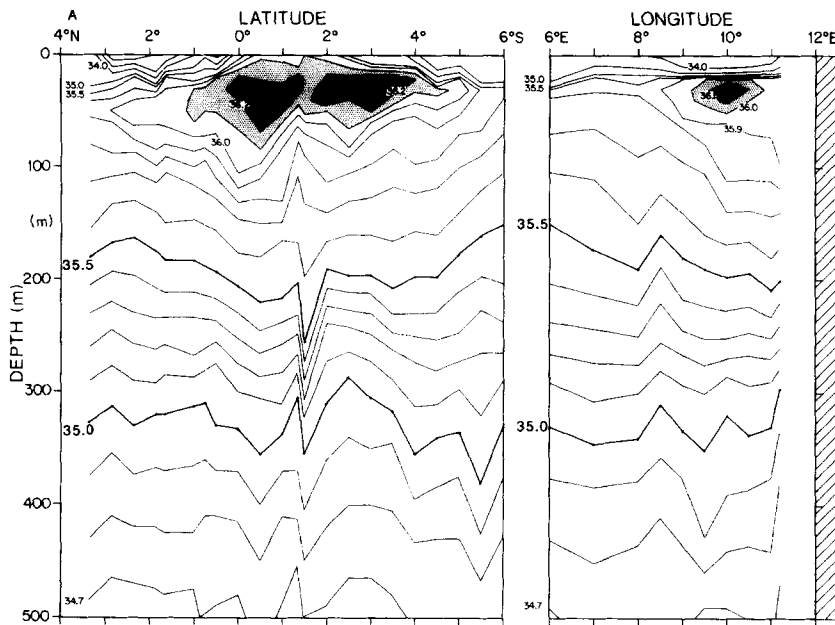


FIG.15a. Salinity sections (0-500m) between $3^{\circ}30'N$ and $6^{\circ}S$ along the Sao Tomé section, and between $6^{\circ}E$ and $11^{\circ}E$ off Pointe Noire, in May 1986 (see Fig.4 for positions). Note the equatorial frontal zones separating fresh waters at $2^{\circ} - 3^{\circ}N$ and $4^{\circ} - 6^{\circ}S$ from the high salinity waters between $1^{\circ}N$ and $4^{\circ}S$.

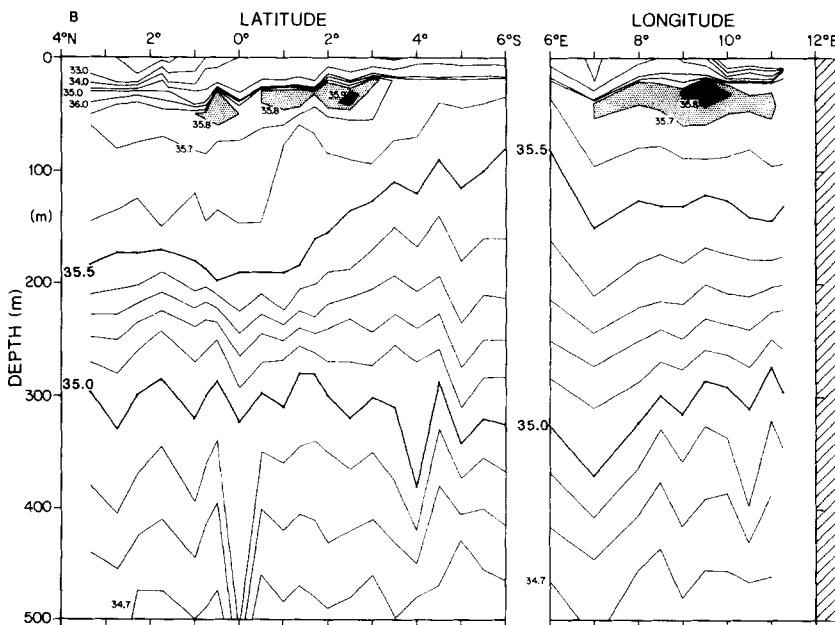


FIG.15b. Same as Fig.15a, in December 1986. Now, the equatorial salinity maximum is strongly reduced and the southern frontal zone has disappeared. Off Pointe Noire, the high salinity core is weakening and spreading longitudinally.

preceding the surface cooling in July-August, is consistent with the theory of McCREARY (1981, 1984). According to the theory of equatorial beams, the upward propagation of the phase of the pressure, density and velocity perturbations is the result of an equatorially-trapped Kelvin wave which propagates energy from changes in the wind stress downwards into the ocean.

4.3 The seasonal cycle along the coast

Along the coast of western Africa in July-September (JAS), distinct cooling occurs only between Lobito (Angola) and Tabou (Ivory Coast). This cycle is weaker at Lagos and vanishes at Conakry (Fig.16). Daily mean sea level at Abidjan and dynamic heights off Abidjan/Grand-Bassam (1966-1971) indicate that the summer time decrease in sea level occurred rapidly at the beginning of June and the minima were observed between July and August (Fig.17). This drop of 10-15cm in sea level, observed all around the Gulf of Guinea in July-September is predominantly steric in origin (VERSTRAETE, 1982, 1985a,b) and cannot be related to the weak seasonal changes of the barometric pressure in the whole area (Fig.1a,b). It is important to note that the minima in sea level occur about one month before the minima in SST, and are not associated with an acceleration of the coastal winds (VERSTRAETE *et al.*, 1980; VERSTRAETE, 1982; VERSTRAETE and PICAUT, 1983). On the other hand, most of the seasonal or interannual changes in coastal sea level result from isostatic adjustment and are not the consequence of changes in the intensity of the very shallow coastal currents. Detailed study of the cooling of the subthermocline waters off Abidjan (5°N) and Pointe-Noire (5°S) shows that in the upper 500m the deep cooling is related to the vertical spreading of the SACW between 500 and 50m (VERSTRAETE, 1985a,b, 1987).

The annual cycle of isotherms and isohalines (0-500m) off Pointe-Noire and Abidjan shows a strong annual upwelling which is related to a distinct salinity maximum within the thermocline (Fig.18a,b). This event at Pointe-Noire precedes that at Abidjan by about one month, with two warm seasons in February-March and October-November (high sea levels) and two cold seasons in December-January and June-July (low sea levels). In both areas, the bottom of the thermocline (14°C isotherm) remains at about 200m throughout the year, and it is the top of the layer (16°C isotherm) which exhibits large vertical moves. The depression of all the isotherms above 14°C, results in a thinning of the thermocline in April at Pointe-Noire and May at Abidjan. The thermocline is best developed during the JAS period. In contrast with Abidjan, the large increase in thermocline thickness off Pointe-Noire results from the strong uplifting of the isotherms above 14°C (at about 200m), without a large change in the depth of the base of the thermocline. Monthly T/S diagrams show that the strong rise of nearly 80m of the 16° - 21°C isotherms between April and August corresponds to a strong intrusion of SACW between 20m and 275m. There is a salinity maximum (35.7 - 35.9) within the rising thermocline off Pointe-Noire in March-July, and within the sinking thermocline off Abidjan in September-November (Fig.18).

4.4 The thermocline during 1982-1984

Near the equator, the thermocline is a layer of almost homogeneous water which is not formed in direct contact with the atmosphere. The thermocline is apparent in the wide spread of isotherms (and isopycnals) just below the tropical thermocline. Thermoclines are a key feature of the eastern equatorial Pacific and Atlantic Oceans, because the waters transported eastwards by the South Equatorial Countercurrent (SECC), the Equatorial Undercurrent (EUC) and the North Equatorial Countercurrent (NECC) must return to the west. At 4°W, two return flows occur on either side of the EUC at 2°-3°N and 2°-3°S (Fig.12a), the northern flow seeming to be the stronger. These

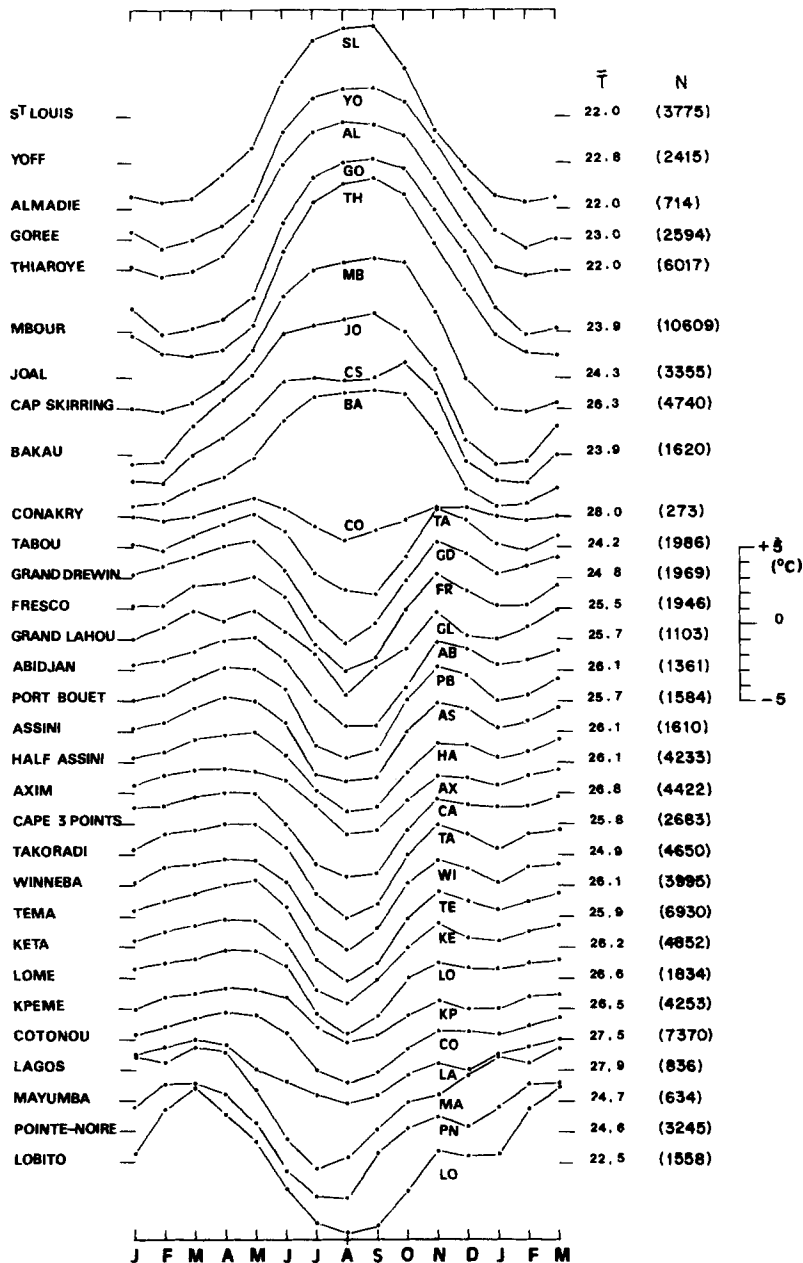


FIG.16. Annual cycle of the SST at 31 West Africa coastal stations from Lobito (12°S) to St Louis (16°N). $\bar{T}$ is the mean temperature of all the daily observations N. The upwelling signal in July-September disappears between Conakry and Bakau. Latitudes of some stations: Gorée, 14°40'N; Bakau, 13°28'N; Conakry, 9°31'N; Abidjan, 5°14'N; Tema, 5°38'N; Lomé, 6°08'N; Cotonou, 6°21'N; Lagos, 6°28'N; Pointe Noire, 4°48'S.

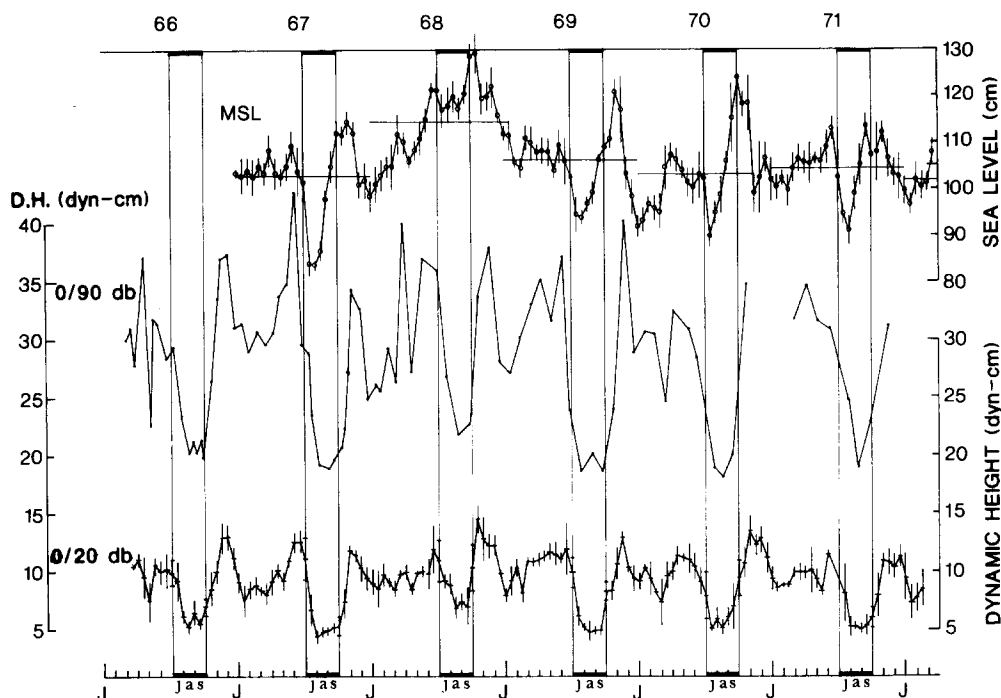


FIG.17. The annual cycle of the observed sea level at the Abidjan-Vridi tide-gauge (MSL in cm) and the dynamic heights (DH in dyn-cm) over the Ivoirian shelf off Abidjan-Grand Bassam from 1966 to 1971. Station-depths: 25m and 200m. Reference level of the sea surface dynamic heights: 20 and 90dbar. One standard deviation is shown for MSL and DH at the Abidjan coastal station situated at 5°10'N, 4°03'W (depth: 25m). The main upwelling season in July-August-September (jas) is underlined. The annual MSL is indicated by a thin horizontal line. Due to unusual heavy rainfalls in 1968, the Vridi tide-gauge did not record the low sea levels related with the seasonal upwelling.

The tide-gauge is situated in the Vridi canal, connecting the lagoon to the ocean.

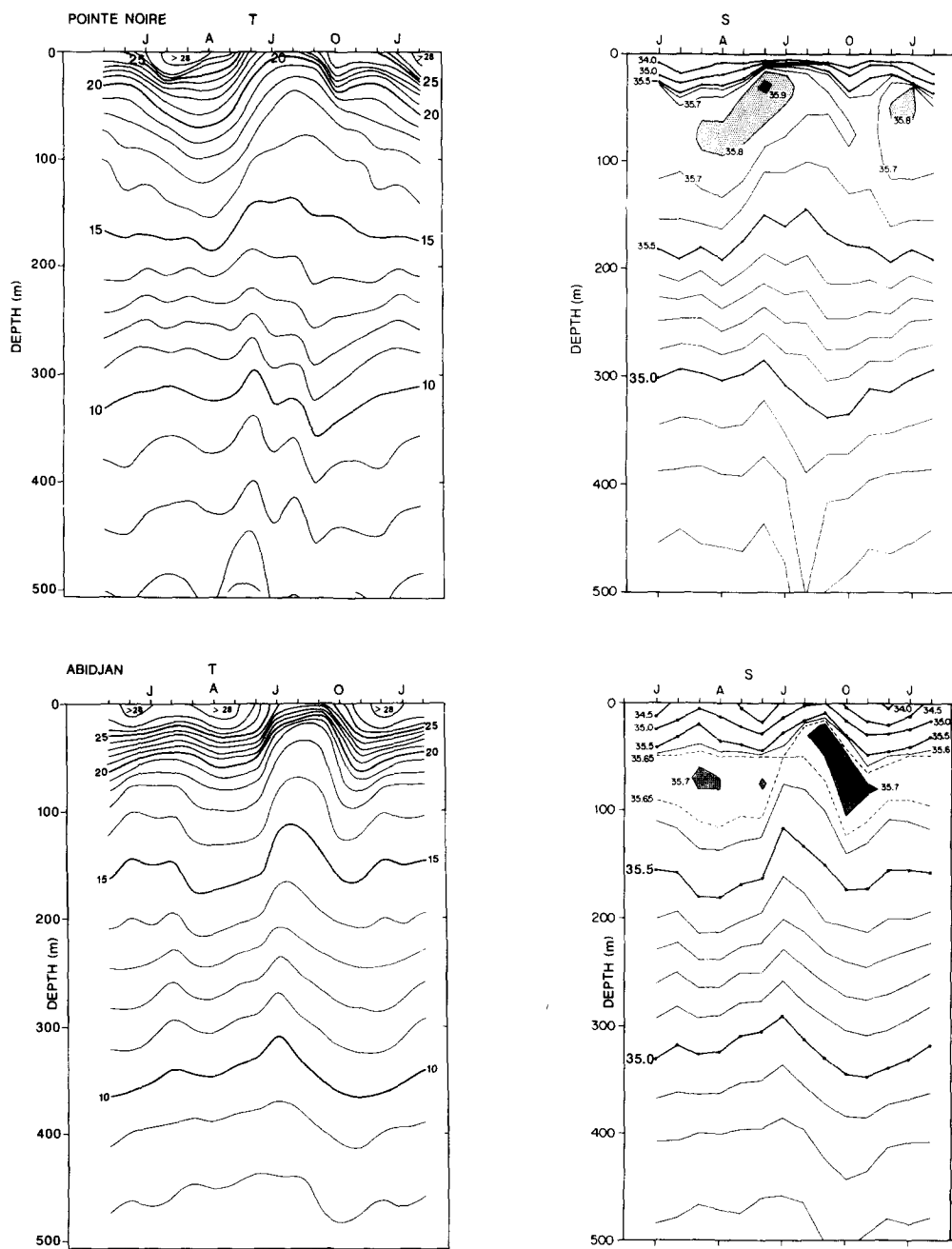


FIG.18. The annual cycles of the temperature (left) and salinity (right) in the upper 500m off Pointe Noire and Abidjan. Note that the thermostat nearly disappears off Pointe Noire in April-May and is reduced in June-July off Abidjan. It is best developed in July-August-September. In the two areas, a distinct cooling and lower salinities appear first at 300-500m in April-May-June.

westward return flows coincide with thermostads which are thicker than at the equator as seen in the mean temperature sections at 4°W, 1°E and 6°E during 1982/1984 (Fig.10). At 10°W eastwards to 6°E the temperature-salinity (T/S) characteristics at the mid-depth of the thermostad coincide with the T/S curves for 170-290m near the equator at 35°W - 29°W. Thus waters with SACW characteristics are carried in the EUC and the SECC from the western-central equatorial Atlantic toward the thermostad regions in the Gulf of Guinea; the thermostad water is moving both eastwards (in the subthermocline part of the EUC and SEUC) and westwards (in the compensation flows). TSUCHIYA (1981, 1986) stipulated that geostrophy requires an equatorial thermostad to develop, with maximum thickness to the North and South of the EUC.

To identify the core of the thermostad in the equatorial Atlantic, we have used the mean equatorial section (35°W - 6°E) and the mean sections at 35°W, 4°W, 1°E and 6°E (Figs 3, 10 and 11), derived from the eight FOCAL cruise data (1982/84). The nearly homogeneous subsurface waters show a regular shoaling and temperature increase from 35°W (11°, 12°, 13°C isotherms; 170-290m) to 6°E (14°, 15°, 16°C isotherms; 110-210m). At 23°W and 4°W, the thermostad is delimited by the 12° and 14°C isotherms (150-260m) and the 13° and 15°C isotherms (115-240m) respectively. KATZ, BRUCE and PETRIE (1980) observed also a warming of the thermostad between 33°W and 10°W. The mean thermosteric surfaces for 1.4, 1.5, 1.6 and 1.8 ml kg⁻¹ were derived from the FOCAL sections at 4°W, 1°E and 6°E. Grids of the thermosteric anomaly 10m vertical and 1/2° latitudinal were computed from which depths of selected isanosteres were linearly interpolated. The arithmetic mean depth (and standard deviation) of each isanostere have then been computed.

Each map reveals zonally organized structures, the depressions (ridges) of isopycnals implying higher (lower) pressure areas. The four maps, corresponding to the levels of the 12°, 13°, 14° and 16°C isotherms, indicated a general deepening of the isanosteres from 5°S, 4°W (relative low pressure area) toward the Sao Tomé-Annobom area (relative higher pressure area). On the two deepest isanosteric surfaces (1.4, 1.5 ml kg⁻¹) two zonal troughs appear, one very well delineated at about 2°S, and a broader one between 1°N and 4°N (Figs 19a,b). The equatorial region 1°S - 1°N appears as a relatively low pressure zone (convergence), with a ridge slightly shifted to 0°30'S on the deeper surfaces. In contrast, the shallower isanosteric surface (1.8 ml kg⁻¹) indicates a depression (high pressure zone) along the equator to the west of Sao Tomé island and in the North-East corner of the Gulf (delineated by the 110m isoline) and two ridges (low pressure zones); the northern one of which is zonally oriented at 1°-2°N whereas the southern ridge curves southward between 5°W and 6°E (Fig.19c). This structure is characteristic of a divergent eastward flow at the equator.

It seems appropriate to delimit the thermostad in the Gulf of Guinea by using the 1.5 (bottom) and 1.8 (top) ml kg⁻¹ isanosteric surfaces. The 1.5 ml kg⁻¹ surface corresponds to temperatures of 13.18°C and salinities of 35.20, and the 1.8 ml kg⁻¹ surface to 16.04°C and 35.65. The resulting map of mean thermostad thickness (Fig.19d), defined by the vertical separation of these two isanosteres, shows two nearly symmetric zonal structures between 4°W and 6°E at about 1°-2°N and 1°-2°S where the thermostad thickness exceeds 160m. The map confirms TSUCHIYA's (1986) results that thermostad thickness tends to be greater to the south of the equator than to the north. Slightly south of the equator (about 0°30'S), the eastward EUC and vertical spreading of the isotherms result in the reduction of the thermostad thickness to 120-150m, and this equatorial minimum extends westwards to the vicinity of Sao Tomé island (Fig.19d).

These zones of high pressures centred at 2°N and 2°S, flanking the low pressure equatorial zone in the 120-270m layer at 4°W, implies subsurface westward flows at 2°N and 2°S and convergence of water toward the equator. Direct observations as well as geostrophic computations at 4°W (both

referenced to 500dbar), confirm the occurrence of two westward flows between 1°30'N(S) and 3°N(S). The mean zonal component of the geostrophic current was computed between 1°N(S) to 5°N(S) from the mean pressure gradient estimated on the 1/2° latitude grid (Fig.20). At the equator, the meridionally differentiated form of the geostrophic balance, $\beta u = (-1/\rho)p_{yy}$, gives unrealistic currents (β is the derivative of the planetary vorticity, ρ the density, p_{yy} the second derivative of the pressure field). This happens because northward winds maintain a southward pressure force in the upper layers. For this reason, the equatorial currents are ageostrophic above the strong equatorial thermocline lying at about 50m in the Gulf. Nevertheless, at 2°N and 2°S, the observed westward flows appear in geostrophic balance, consistent with the maximum development of the thermostad there (TSUCHIYA, 1981, 1986).

At 6°E, PCM observations show a westward flow at about 1°30'S in the layer 50-200m (Fig.12b). These results show the dynamical relationship between the EUC, the large thermostads along its flanks and the deep westward geostrophic compensation flows at 2°N and 2°S.

Polewards, the thermostad thickness is reduced, more dramatically south of 3°S than to the north. Direct observations and geostrophic currents are in remarkable agreement at 4°-5°S between 80 and 200m (Figs 12a, 20). The distribution of the mean thermostad thickness along 5°S as well as the meridional sea surface dynamic topography at 6°E suggest that the trajectory of the SECC (surface and subsurface) bends southeastward so that it flows nearly parallel to the African coast.

In summary, the general pattern that emerges from these analyses is one of a divergence of near surface waters (temperatures above 16°C) at the equator, deepening from 80m at 4°W to 120m at 6°E (Fig.19c) and convergence of subsurface waters (80-270m at 4°W; 120-270m at 6°E, Fig.19a,c) towards and along the equator as in the schematic diagram of the currents of PHILANDER and PACANOWSKI (1986b). Two westward subsurface flows flank the EUC at about 2°N and 2°S and appear in geostrophic balance. These results are in agreement with TSUCHIYA (1986) and suggest that horizontal convergence in the lower portion and below the EUC is basically responsible for the existence of the thermostad. It is unlikely that vertical mixing can create a thermostad in a strongly stratified ocean where at best this process could help the seasonal increase of the thermostad. Analysis of the Richardson number by GOURIOU (1990) gives evidence of a larger turbulence at the equator and above the EUC than off the equator and below the EUC. During 1982-84, in the Gulf of Guinea, the average flow in the core of the EUC, estimated from eight campaigns and within the 10cm s⁻¹ isotach, was divergent and decreasing from about 9 Sverdrups at 4°W to 2.7 Sverdrups at 6°E (standard deviations are 3.9 Sverdrups and 2.1 Sverdrups, respectively).

4.5 Dynamic topography and geostrophic flow during 1982 - 1984 (ref.500dbar)

The mean dynamic surface topography shows a ridge at about 2°-3°N flanked by low dynamic heights (Fig.21a). This ridge is a consequence of the strong equatorial front that runs latitudinally at about 2°N in the Gulf of Guinea, separating the warm and low salinity (26°-28°C, 34.5-35.1) tropical surface water north of 2°N from the relatively colder and saltier equatorial waters. The low dynamic heights along the northern African coast are the signature of the strong and shallow geostrophic Guinea Current flowing eastward between 3°N and the coast (Figs 12a,20). At about 2°N, a strong westward geostrophic flow is identified as the South Equatorial Current (SEC). At 1°S, the surface geostrophic flow is eastward and opposite to the surface Ekman drift associated with the southerly winds. Note that along the northern coast at 4°-5°N, the sea surface topography slopes upwards from 4°W to 6°E (Fig.21a). This westward zonal pressure gradient is responsible for the coastal Guinea undercurrent (Fig.20).

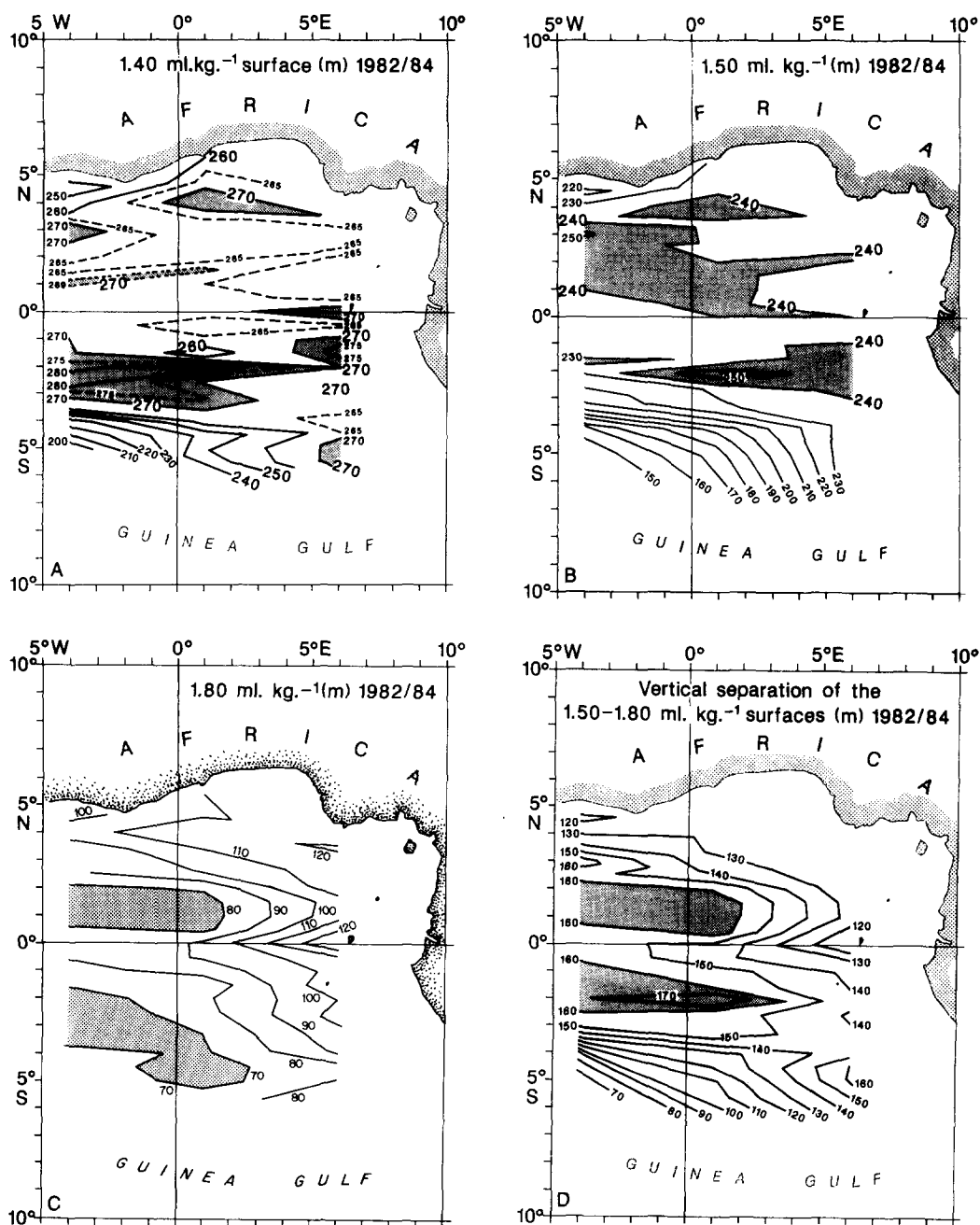


FIG. 19. Maps of the mean depth of the isanosteric surfaces in the Gulf of Guinea during 1982-1984 derived from the FOCAL sections at 4°W, 1°E and 6°E. (a) 1.4 ml kg.⁻¹ surface, depressions are stippled (higher pressure areas); (b) 1.5 ml kg.⁻¹ surface, depressions are stippled; (c) 1.8 ml kg.⁻¹ surface, the two ridges (low pressure zone) are stippled; (d) mean thickness of the thermocline as defined by the vertical separation of the 1.5 (bottom) and 1.8 ml kg.⁻¹ (top) surfaces. Areas where the thickness exceeds 160 m are stippled.

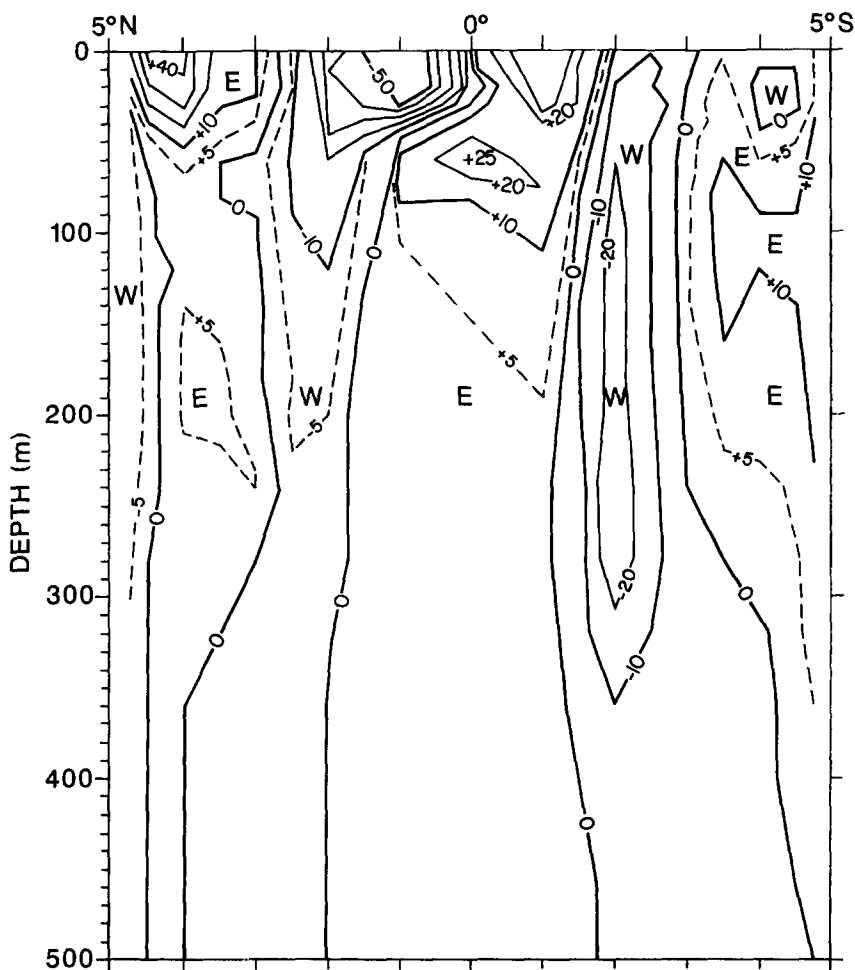


FIG.20. Meridional section of mean zonal geostrophic current at 4°W on a $1/2^{\circ}$ latitudinal grid, between 5°N and 5°S . At the equator, the meridional derivative of the geostrophic balance was used and the computed currents are not realistic between 1°N and 1°S . Note the two westward geostrophic flows at $2^{\circ} - 3^{\circ}\text{N,S}$; the Guinea current at $3^{\circ} - 5^{\circ}\text{N}$ and the westward coastal Guinea under current. The SEUC flows eastwards at $3^{\circ} - 5^{\circ}\text{S}$ below 50m.

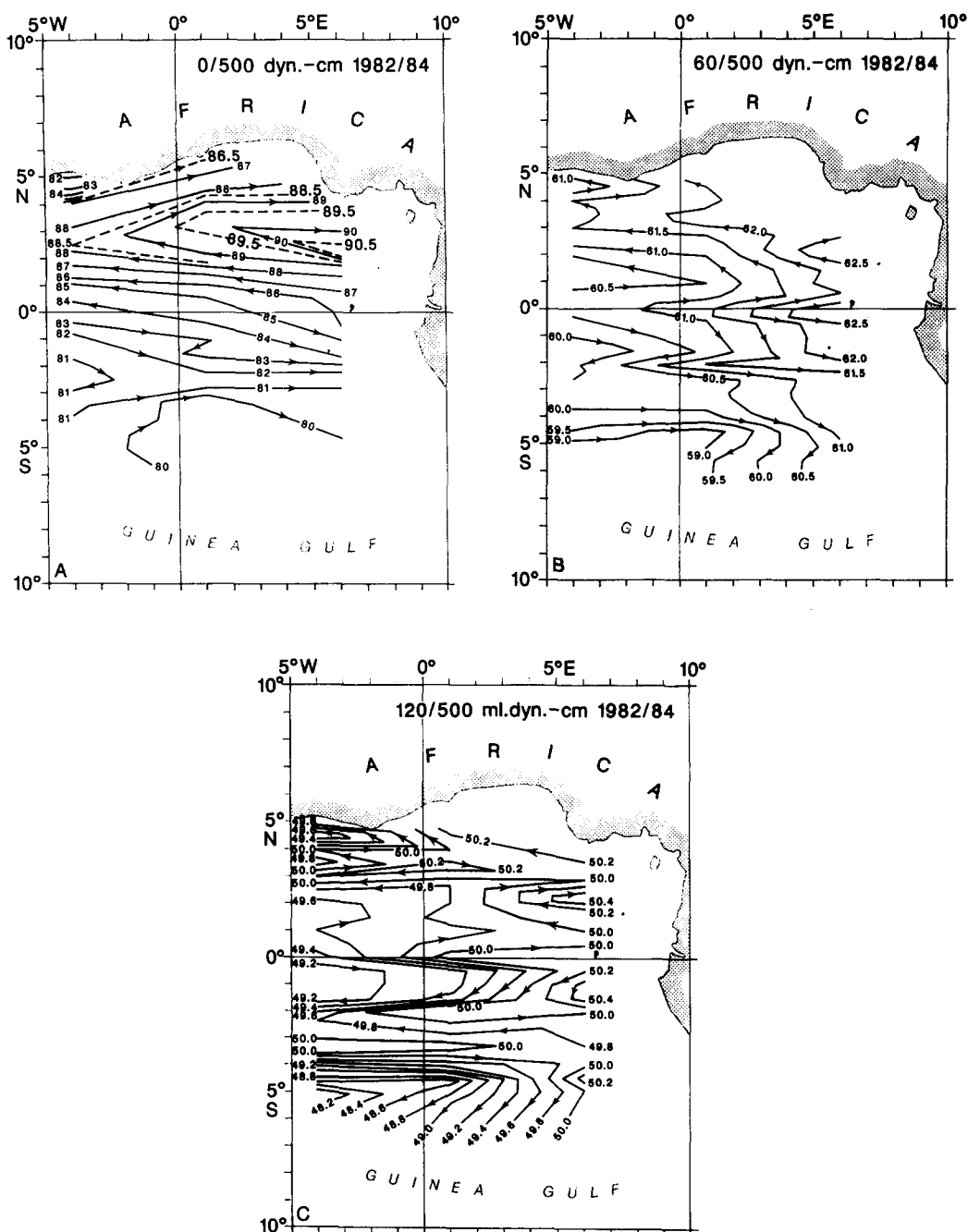


FIG.21. Mean dynamic topography referenced to 500dbar in the Gulf of Guinea during 1982-1984, derived from the FOCAL sections at 4°W, 1°E and 6°E. (a) Sea surface, contour interval is 1 dyn cm⁻¹; (b) 60dbar surface, contour interval is 0.5 dyn cm⁻¹; (c) 120dbar surface, contour interval is 0.2 dyn cm⁻¹

At 60dbar (Fig.21b), depth of the EUC, the flow is eastward between 1°N and 2°S and southeastward south of 3°S and to the east of the first meridian. This southeastward flow at 6°E is probably related to the subsurface flow observed over the shelf break off Africa between $1^{\circ}30'\text{S}$ and 6°S (WACONGNE, 1988). Note the two geostrophic westward compensation flows at 2°N and 2°S flanking the EUC (Figs 12a,20), and the westward coastal Guinea under current at 5°N .

Finally, the 120dbar surface (Fig.21c) reveals the main features of the subthermocline circulation in the Gulf of Guinea. From north to south we note: (1) the westward flowing subthermocline Guinea Countercurrent with its two components, the coastal Guinea undercurrent flowing along the shelf break at about 60-80m and the Guinea countercurrent at 2° - 3°N which is the northern part of the return flow of the EUC; (2) the EUC which from direct observations has speeds greater than 15cm s^{-1} at the equator. Here the application of geostrophy seems better because at 120dbar the currents are isolated from the weak cross-equatorial winds by the strong thermocline (HOUGHTON, 1989; VERSTRAETE and VASSIE, 1990); (3) at 2° - 3°S , the second compensation westward geostrophic flow and (4) the subthermocline eastward South Equatorial Undercurrent (SEUC) flowing between 4° and 5°S between 50 and 300m. There is also a south-westward flow at 5° - 6°S , outside the thermostat region, which is probably related to the return flow of the SEUC (TSUCHIYA, 1986). The flows (1) - (4) are clearly seen at 4°W , both on the zonal and the geostrophic zonal components of the observations (Figs 12a,20). The dynamic topography at 60db and 120db shows clearly that the eastward flow along the equator is divergent (Fig.21b,c).

These analyses are consistent with the direct current observations at 6°E , limited to the band

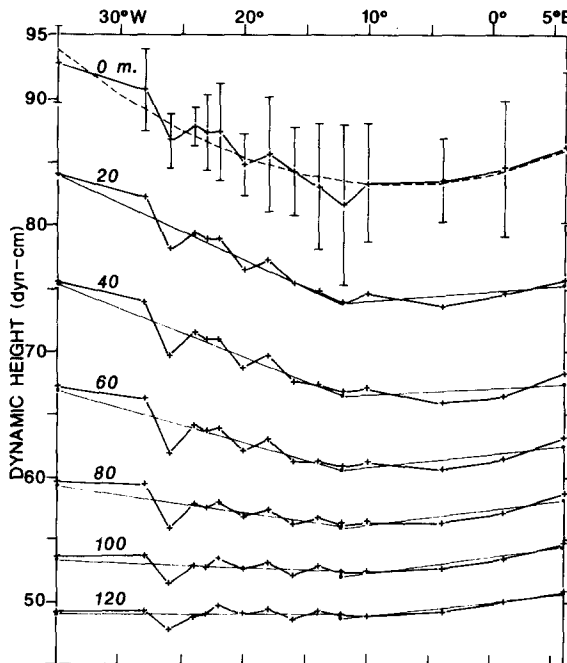


FIG.22. Mean dynamic height (reference 500dbar) from the surface to 120dbar by steps of 20dbar along the equator between 35°W and 6°E (mean of the 8 FOCAL campaigns from October 1982 to September 1984). Units are dyn cm. Dashed line: surface parabolic adjustment (89% of the variance explained). The 95% confidence intervals are indicated at the sea surface. The linear fits are computed from 11 stations (5 stations) between 35°W and 12°W (12°W and 6°E). Note the reversal of the sea surface slope at about 10°W .

1°30'N - 3°30'S (Fig.12b). Finally, Fig.21 fails to depict adequately the reversal of the zonal dynamic topography at about 10°W along the equator in the Gulf (Fig.22). As we shall see it, the variations of the zonal pressure gradient is one of the keys to understand the mechanism of the seasonal equatorial downwelling-upwelling in the Guinea Gulf.

5. THE DOWNWELLING-UPWELLING PROCESS IN THE EASTERN EQUATORIAL ATLANTIC DURING 1984

5.1 *The mean zonal pressure gradients along the equator between 35°W and 6°E*

The mean depth of the 20°C isotherm, representative of the thermocline, and the mean sea surface dynamic height (ref.500dbar) between 35°W and 6°E, computed from two complete seasonal cycles in 1982-84, were described by VERSTRAETE and VASSIE (1990). These agree generally with the results given by HISARD and HÉNIN (1987) and historical data over the whole equatorial Atlantic (MERLE and ARNAULT, 1985). Mean dynamic heights (ref.500dbar) from the surface to 120dbar by steps of 20dbar along the equator between 35°W and 6°E are shown on Fig.22. Both to the west and to the east of about 12°W, the sea surface sloped upwards toward South America and towards Africa respectively, reflecting the thermal structure in the upper 300m. Whereas the thermocline sloped downwards to the West of the area 0°-5°W and was nearly horizontal in the Gulf of Guinea, the thermocline was warmer at 6°E than at 4°W and 35°W (Fig.11a).

In the Central and Western equatorial Atlantic (35° - 12°W), the eastward integrated mean ZPG in the layer 0-120m was in equilibrium with the mean westward component of the wind stress in the band 2°N - 2°S. A correlation between vertically integrated pressure gradients (to 120m) and the westward wind stress component for the eight FOCAL cruises was highly significant at the 0.01 level with a correlation coefficient of 0.97. However, this result is misleading because an inspection of the individual integrated ZPG (0-120m) and coincident zonal wind stress τ_x (2°N - 2°S) shows that both in January-February 1984 and in April-May 1984, the two fields were far from being in equilibrium.

In the eastern equatorial Atlantic (12°W - 6°E), the westward integrated mean ZPG reached about 200m and was not in equilibrium with the average zonal wind stress component which was eastward only in a narrow band at 4°E - 8°E, and was always very weak. A study of the Sverdrup balance in the Gulf of Guinea in 1982-84 showed that the frictional influence of the average local meridional wind stress component was unable to penetrate through the strong tropical thermocline lying at 50m (VERSTRAETE and VASSIE, 1990). HOUGHTON (1989) arrived at the same conclusion from the analysis of the 4°W equatorial mooring data.

5.2 *The wind forcing during 1983-1984*

By chance during the program of systematic observations in 1983-84, we observed an exceptional warm event in early 1984 in the Gulf of Guinea. Following a relaxation of the trade winds (TW) and southerly monsoon in December 1983 - January 1984, the winds collapsed nearly completely all along the equator in January-April 1984 (Figs 23,24,25). In November 1983 the tropical thermocline was at about 50m, near its climatic depth, but in February 1984 it was observed to be nearly 50m deeper than climatology (Fig.24c). At the same time the sea levels at Principe-Sao Tomé-Annobom islands rose, the surge reaching about 13-15cm at 6°E (Fig.25a) and simultaneously, the eastward ZPG in the western-central Atlantic nearly disappeared (KATZ,

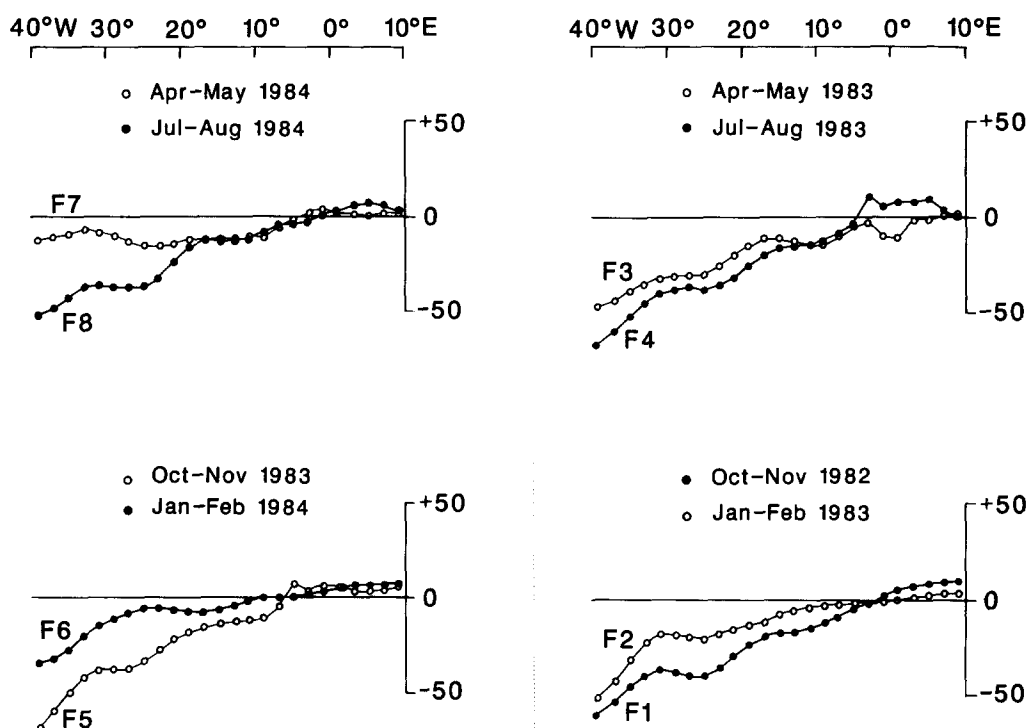


FIG.23. Zonal wind stress distribution (pseudo-stress in m^2s^{-2}) along the equator between 2°N and 2°S , and from 10°E to 40°W during each of the eight FOCAL campaigns (F1-F8). Note the collapse of the zonal wind stress all along the equator in January-February 1984 (F6) and in April-May 1984 (F7).

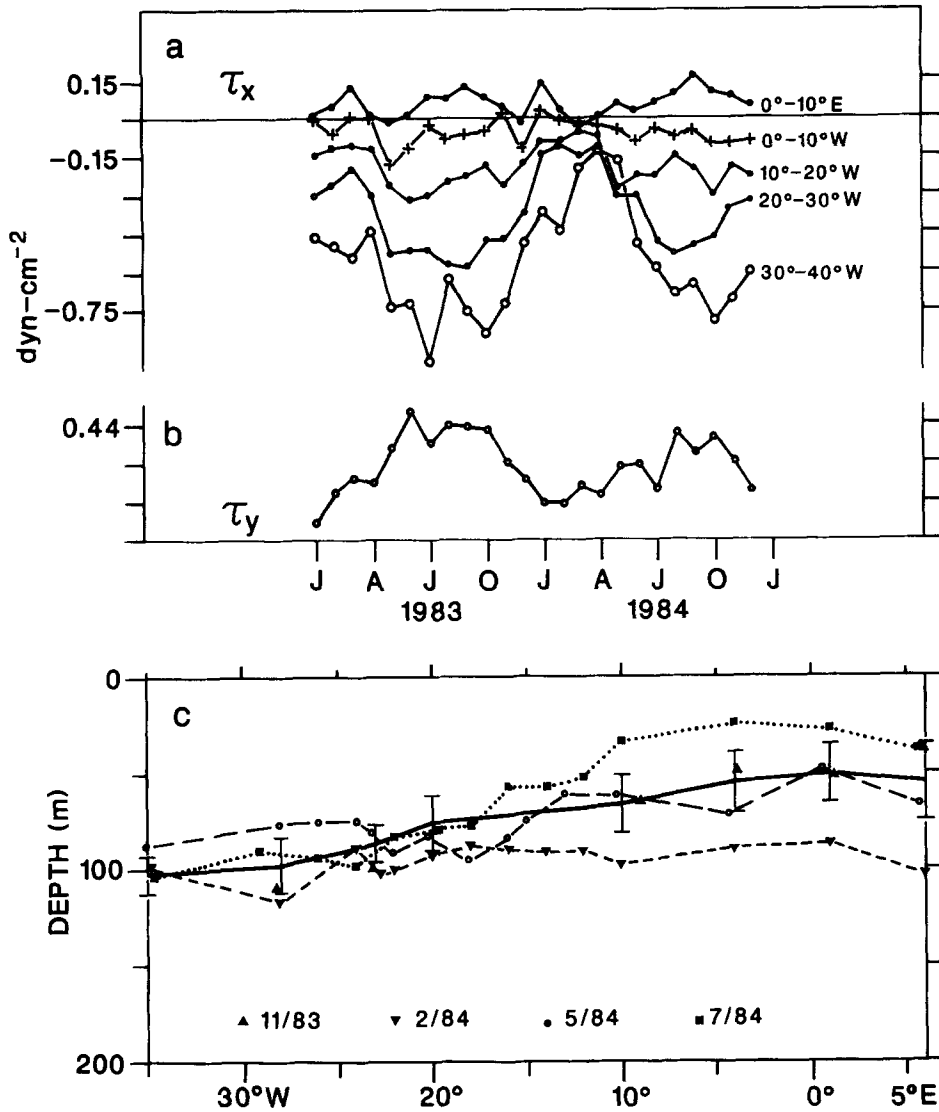


FIG.24 (a) The monthly mean zonal wind stress τ_x in 10° wide bands along the equatorial Atlantic during 1983 and 1984, between 10°E and 40°W . (b) In the Gulf of Guinea, the monthly mean meridional wind stress τ_y is given in the area $6^\circ\text{S}-6^\circ\text{N}, 6^\circ\text{E}-8^\circ\text{E}$. (c) the mean topography of 20°C isotherm depth (nearly the depth of the tropical thermocline) between 35°W and 6°E during 1982-1984; the 95% confidence intervals are indicated. Its vertical displacements are shown from November 1983 to July 1984.

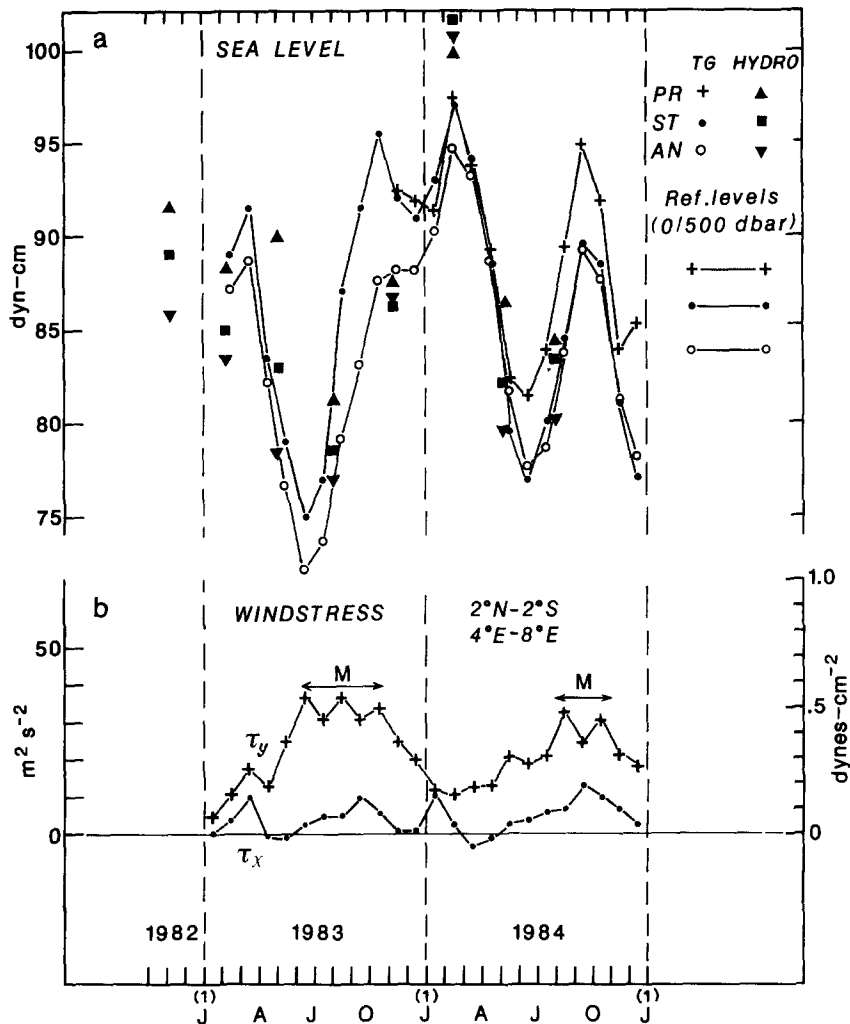


FIG.25 (a) Monthly mean sea levels at Principe (PR), Sao Tomé (ST) and Annobom (AN) during 1983 and 1984. Each tide-gauge (TG) is calibrated against the mean dynamic height (Ref.levels) computed from the observed hydrology in 1982-1984: (HYDROlogy). Sea levels are in cm, dynamic heights in dyn-cm. (b) Monthly wind stress (τ_x zonal and τ_y meridional) in the area of the islands during 1983 and 1984. The period of the African monsoon is labelled "M". Units are in $m^2 s^{-2}$ (pseudo stress) and in dynes cm^{-2} (stress, computed with a drag coefficient of $1.2 \cdot 10^3$).

HISARD, VERSTRAETE and GARZOLI, 1986; VERSTRAETE and VASSIE, 1990).

The following annual spring increase of the zonal component τ_x of the TW occurred in April-May 1984 in the equatorial strip $10^\circ-30^\circ W$ (Fig.24a). Data from the meteorological station at RSPP showed that the onset of the TW occurred later there, not before the third week of May 1984 (KATZ, 1987a). During March-April-May 1984, the surge at $6^\circ E$ collapsed with a sea level dropping of about 12cm. It is interesting to observe that the minimum in sea level occurred in June-July 1984, before the onset of the African monsoon, which was fully developed only in August-

October 1984 (Fig.25a,b). In contradiction with a local forcing, the depth of the 20°C isotherm along 6°E, 4°W and 23°W was not significantly deeper in July-August 1983 than in July-August 1984 (HISARD, HÉNIN, HOUGHTON, PITON and RUAL, 1986) although the African monsoon (Fig.25b) was normal in its intensity and timing in 1983 (onset in April-May) but was weaker and late in 1984 (onset in July-August).

The timing of the deepening of the equatorial thermocline in the Gulf in February 1984 was totally unrelated to the local forcing. Moreover, the superficial equatorial seasonal upwelling was inhibited, as SST was about 3°C warmer in June-August 1984 than over the same period in 1983, both at 6°E (T/P gauge, VERSTRAETE, 1988) and at the 4°W equatorial mooring (WEISBERG and COLIN, 1986). If the local southerly monsoon winds had been forcing an equatorial Ekman divergence and so driving the thermal structure, we should have had observed cold SST and low sea levels from June (August) to October in 1983 (1984). In reality, we observed minima in June-July both in 1983 and 1984 (Fig.25) which were then followed by rising sea level and SST from July to September-October, indicating that neither the intensity nor the timing of the equatorial downwelling-upwelling in the Gulf of Guinea were correlated with the local wind forcing during 1983-84. The numerous studies which followed the FINE workshop thirteen years ago (MOORE *et al*, 1978; O'BRIEN *et al*, 1978), now enable us to relate changes in the southeast TW in the central-western equatorial Atlantic with basin-wide changes in zonal thermocline depth, zonal pressure gradient (ZPG) and equatorial downwelling-upwelling in the Gulf of Guinea.

5.3 The downwelling-upwelling process in the Gulf in 1984

5.3.1 Equatorial sea surface dynamic topography. Based on linear and parabolic fits, the changes in the sea surface dynamic topography along the equator, together with the vertical structure of the ZPGs from 0 to 120m by 20m steps, from November 1982 to August 1984, are shown in Fig.26. The 95% confidence intervals of the ZPGs is indicated for each linear fit when the number of stations n is greater than 4 (Fig.26).

The downwelling event observed in January-February 1984 in the Gulf of Guinea followed the basin-wide wind relaxation of December 1983 - January 1984. Between November 1983 and February 1984, we observed a strong downwelling at both 4°W and 6°E, where the 19°C isotherm deepened 38m and 90m, respectively. At the same time the equatorial sea surface dynamic topography (Fig.26e,f) changed from a typical profile, to one in which the topography was nearly flat to the west of 15°W and sloped steeply upwards between 10°W and 6°E. In October-November 1983 (Fig.26e), the vertically integrated ZPG to 120m (IZPG) between 10°W and 30°W was nearly in equilibrium with τ_x (mean wind stress computed in the equatorial strip 10°W - 30°W, between 2°N and 2°S).

In January-February 1984 (Fig.26f), a least square fit to a second order polynomial (coefficient of determination $R^2=0.86$) between 35°W and 6°E showed that IZPG was nearly zero at 20°W and positive between 20°W and 6°E with estimates increasing from +0.18 dyne cm⁻² (at 10°W) to 0.43 dyne cm⁻² (at 6°E). However, at the same time, τ_x was negative between 10°W and 35°W (-0.152 dyne cm⁻²) and very weakly positive only between 10°W and 10°E (+0.055 dyne cm⁻²). It was impossible to find an equatorial strip where τ_x and IZPG were in equilibrium in January-February 1984. Synoptic observations at 10°W and 1°W (Inverted Echo sounders) and at 6°E (T/P gauge) showed that at this time the westward ZPG between 6°E and 10°W reached an absolute maximum for the 23 months of coincident observations (VERSTRAETE and VASSIE, 1990). According to the continuous sea level records at and near the equator, the surge in the Gulf at 6°E lasted for nearly three months (January-February-March 1984) and then collapsed in March-

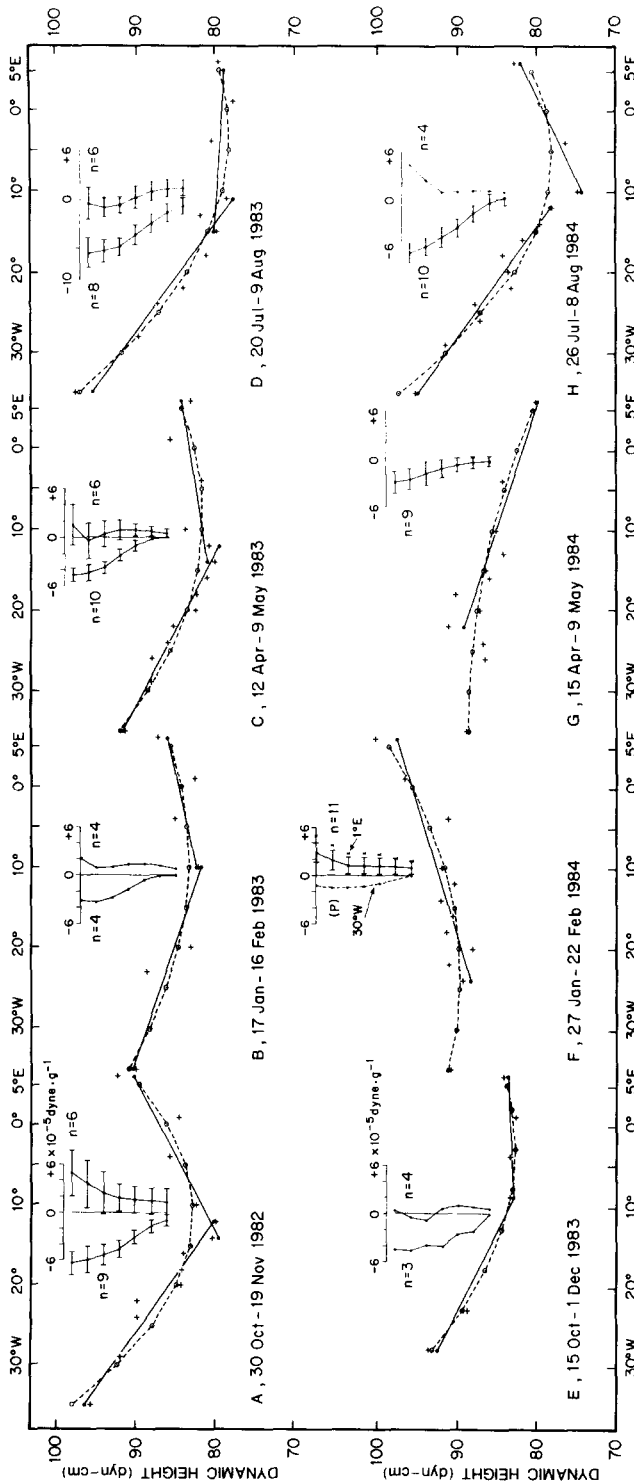


FIG. 26. Sea surface dynamic topography (dyn-cm) of the equatorial Atlantic between 35°W and 6°E during each of the eight FOCAL campaigns: (a) November 1982; (b) February 1983; (c) May 1983; (d) August 1983; (e) November 1983; (f) February 1984; (g) May 1984; (h) August 1984. The best fits (linear and parabolic) are shown together with the computed dynamic heights (ref. 500dbar). Estimates of the zonal pressure gradients in $10^{-5} \text{ dyne g}^{-1}$ from 0 to 120dbar by steps of 20dbar are given for each sea surface topography (with their 95% confidence intervals). Note the change in the vertical and horizontal structure between (e) and (f) (downwelling in the gulf and between (f) and (g) (upwelling). The number of stations (+) n is indicated with each linear fit.

April-May 1984 (Fig.25a).

During the collapse of the winds all along the equator from January to April 1984, we observed a huge swing of the equatorial dynamic topography from February to April-May (Fig.27), but no significant change either in the zonal or meridional wind stress which were extremely weak. Thus it is very unlikely that the ocean was responding to a local wind forcing when the westward Guinea Gulf ZPG reversed direction to become eastwards. On the other hand, the hypothesis of a basin-wide equatorial "seiche" is untenable, because we did not observe a large symmetrical signal at 30-35°W (Fig.27).

In the central equatorial Atlantic (10°-30°W), τ_x increased in April-May 1984, as in the climatology, before the seasonal increase in the western basin (30°-50°W) and at RSPP. The 19°C isotherm at 4°W and 6°E rose 13m and 62m, respectively. From a linear fit between 22°W and 6°E (Fig.26g, $R^2=0.83$ at 0m), a comparison between IZPG ($-0.194 \text{ dyne cm}^{-2}$) with τ_x in the same equatorial strip ($-0.080 \text{ dyne cm}^{-2}$) revealed no equilibrium nor did any other ZPG estimates from a parabolic fit (Fig.27, $R^2=0.69$ at 0m) find any relation between the IZPG and τ_x in various equatorial strips.

Finally, in July-August 1984 the equatorial Atlantic had nearly recovered its usual sea surface topography with an upward sea surface slope from 10°W towards Africa (Fig.26h). In this upwelling situation in the Gulf, the IZPG between 12°W and 35°W ($-0.39 \text{ dyne cm}^{-2}$) was nearly

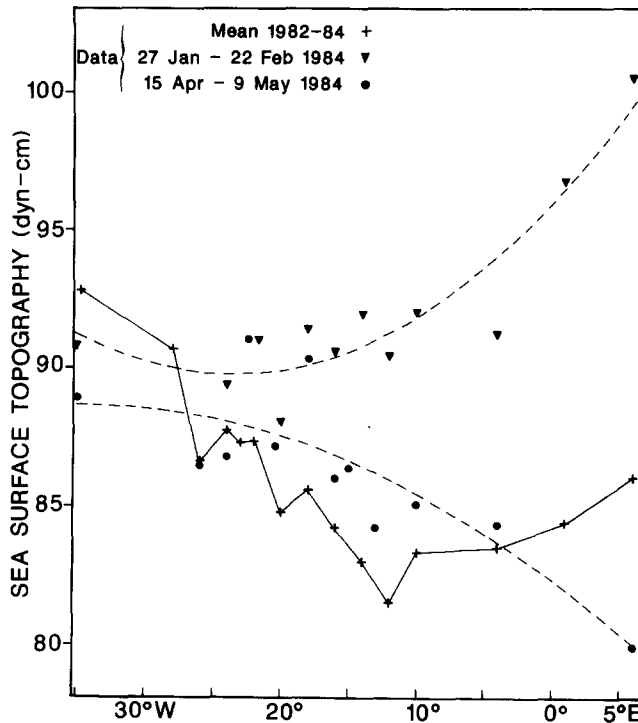


FIG.27. Change in the sea surface topography (ref.500dbar) between January-February 1984 and April-May 1984 together with the 1982-1984 mean. Twelve stations were used for each parabolic fit (dashed lines). The determination coefficient R^2 is .86 in January-February 1984 and .69 in April-May 1984. Units: dyn-cm.

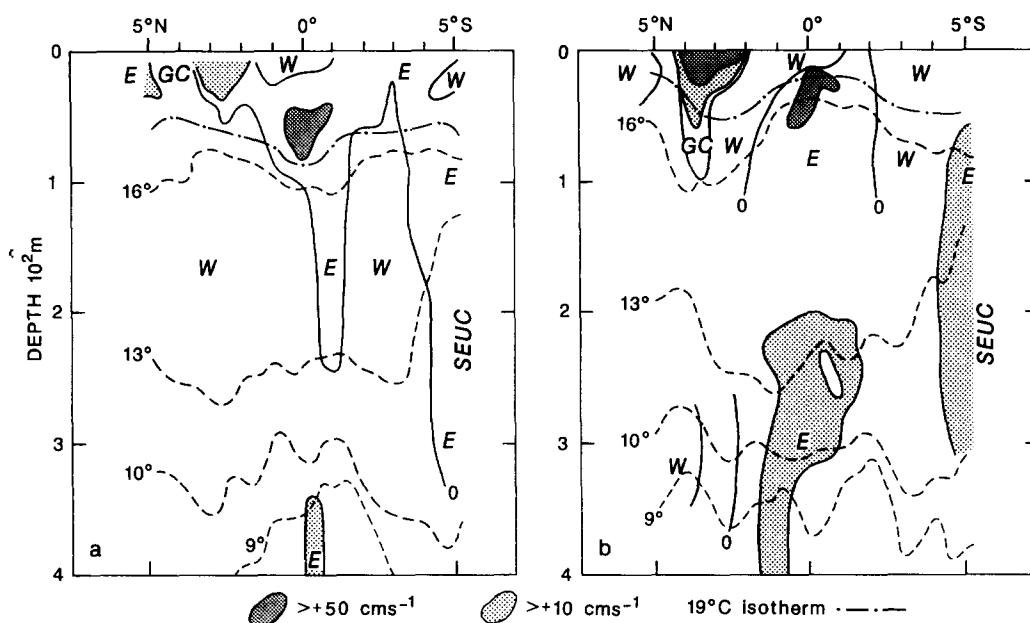


FIG.28. Zonal component of the currents observed at 4°W between 5°N and 5°S in the layer 0-400m. Reference 500m. Superimposed are shown the isotherms 9°, 10°, 13°, 16° and 19°C. Eastward currents (E) are lightly (darkly) stippled above 10cm s⁻¹ (50cm s⁻¹). GC is for Guinea Current, SEUC for South Equatorial Under Current. The 0 isotach separates the Eastward flow from the Westward flow (W). (a) February 1984; note the two westward flows coincident with a very thick thermostat (13° - 16°C) at 2° - 3°N,S; the pinching of the thermostat between 1°N and 1°S. The equatorial under current flows above the 19°C isotherm and the top of the thermostat (see Fig.32a). A deep eastward jet appears at the equator below 350m. GC is weak and shallow. (b) July 1984; note the deep eastward jet at the equator between 200 and 400m (maximum speed greater than 20cm s⁻¹ at 250m and 0°30'S); the deepening and strengthening of GC at 2° - 4°N; now the thermostat is pinched at 4°N and 5°S, very thick between 2°N and 2°S. The equatorial under current flows below the 19°C isotherm and the top of the thermostat (Fig.32a). All the isotherms are strongly uplifted between 3°N and the coast, when only the shallower isotherms are uplifted at the equator. The bottom of the thermostat is practically unchanged whereas the 13°C isotherm slopes steeply at 3° - 4°N.

in equilibrium with τ_x ; but in the Gulf there was again no equilibrium ($\text{IZPG} = +0.187 \text{ dyne cm}^{-2}$ and $\tau_x = +0.008 \text{ dyne cm}^{-2}$). At 4°W and 6°E the 19°C isotherm had risen 52m and 29m respectively.

5.3.2 The zonal currents at 4°W in February and July 1984. In November 1983, the EUC at 4°W had a large vertical extension, the 10cm s⁻¹ isotach reaching 250m. Between November 1983 and February 1984, the depth of the shallow core of the EUC did not change during the downwelling event either at 4°W (core at 60-70m) or at 6°E (60m). However, its flow increased slightly from about 6 to 7 Sverdrups at 4°W and from 2 to 3 Sverdrups at 6°E (estimates by planimetry, contouring the isotachs equal or greater than 10cm s⁻¹). In February 1984, the Guinea current (GC) was weak and shallow (Fig.28a). An unusual eastward 0-20m surface current was observed at 3°-4°S, branched with the South Equatorial Under Current (SEUC), which was flowing eastwards as usual between 50m and 300m at 3°-5°S (Fig.28a). More significantly, the vertical extension of the EUC had been strongly reduced (the 10cm s⁻¹ isotach being not deeper than 100m) and the weak westward flow observed between 300 and 500m in November 1983 had

been replaced by a distinct eastward flow nearly at the equator, which had speeds greater than 10cm s^{-1} at $0^{\circ}30'S$, and was separated from the shallow EUC core (Fig.28a).

In May 1984 (not shown), a deep eastward jet was observed within $1^{\circ}30'$ of the equator, between 250 and 350m, with a maximum speed of 35cm s^{-1} at $0^{\circ}30'S$; the core of this deep jet was lying at 300m and its transport was estimated to be about 3 Sverdrups.

Finally in July 1984 (Fig.28b), the cores of the EUC and of the deep eastward jet had shoaled to about 50m and 250m respectively. The Guinea current was very swift ($+90\text{cm s}^{-1}$) and deep (100m). The speed of the deep jet had reduced to about 20cm s^{-1} but its transport had increased to 4.6 Sverdrups (between $1^{\circ}30'N$ and $1^{\circ}30'S$, 200 and 400m, within the 10cm s^{-1} isotach).

From February to July 1984, it is noteworthy that both the transport (about 7 Sverdrups) and the maximum speed (about 70cm s^{-1}) of the EUC at $4^{\circ}W$ were nearly insensitive to large changes in the thermal structure. However, the core of the EUC was uplifted by about 20m (Fig.28). During the same time period the transport of the deep eastward jet had increased from less than 1 Sverdrup to about 4.6 Sverdrups, comparable to the transport of the EUC. In May (July 1984), this deep jet was lying within (above) the deep equatorial thermocline (approximated by the $10^{\circ}C$ isotherm). The spreading of the 9° - $13^{\circ}C$ isotherms in July 1984 between $1^{\circ}30'N$ and $1^{\circ}30'S$ is indicative of a geostrophic balance: isotherms above $10^{\circ}C$ were elevated but the deeper ones depressed (Fig.28b). Geostrophic computations below 100m show indeed an eastward zonal flow between 250 and 400m and $1^{\circ}N$ and $1^{\circ}S$, with geostrophic speeds greater than 20cm s^{-1} .

5.3.3 Changes in depth of the $19^{\circ}C$ isotherm. The main features of the thermal structure are superimposed on the zonal currents in Fig.28. At $4^{\circ}W$, the $13^{\circ}C$ and $16^{\circ}C$ isotherms delineate the thermocline, approximately (Fig.29a). The $10^{\circ}C$ isotherm visualises the deep permanent thermocline at about 330m (Figs 10 and 11a) whereas the $19^{\circ}C$ isotherm lies in the lower limit of the shallow tropical thermocline (Fig.29). To describe simply the changes in the meridional thermal structure in the shallow tropical thermocline, I have used the changes in depth of the $19^{\circ}C$ isotherm. This choice was dictated by an examination of temperature and zonal current sections which showed that the $19^{\circ}C$ isotherm is normally found in lower part of the thermocline and near the core of the EUC (Figs 11a,28), deep enough to be well sheltered from the near surface wind mixed layer in this region.

During the downwelling event, the changes in depth of the $19^{\circ}C$ isotherm with latitude $\Delta d(19)$, between November 1983 and February 1984 are suggestive of a Gaussian profile (Fig.30). A fit of the form $\Delta d = ae^{by^2}$ as a function of the distance y from the equator was computed using least squares regression. Both at $4^{\circ}W$ and $6^{\circ}E$, the perturbation in the temperature field was nearly symmetrical about, and was maximum at the equator (maximum deepening of 38m at $4^{\circ}W$ and 90m at $6^{\circ}E$, minima at 2° - $3^{\circ}N$, S). In this case, the regression coefficient b is related to the equatorial scales of the problem by $b = 1/2L^2$ where L is the equatorial Rossby radius of deformation (or the trapping scale). The phase speed and equivalent depth are respectively:

$$c = -\beta/2b \quad \text{and} \quad h = c^2/g$$

where β is the planetary vorticity gradient, g the gravitational acceleration. Two fits between $3^{\circ}30'S$ and $3^{\circ}30'N$ at $4^{\circ}W$ and $6^{\circ}E$ explained 88% and 47% of the variance, respectively. At $4^{\circ}W$, the trapping scale was 221km, the speed of the Kelvin wave 112cm s^{-1} and the equivalent depth 13cm. These scales compare favourably with those associated with the second baroclinic mode in the Gulf of Guinea. At $6^{\circ}E$, the fit gave the following results: $L=292\text{km}$, $c=196\text{cm s}^{-1}$ and $h=39\text{cm}$, which compare with values comprised between the first and second baroclinic modes (Tables 1 and 2).

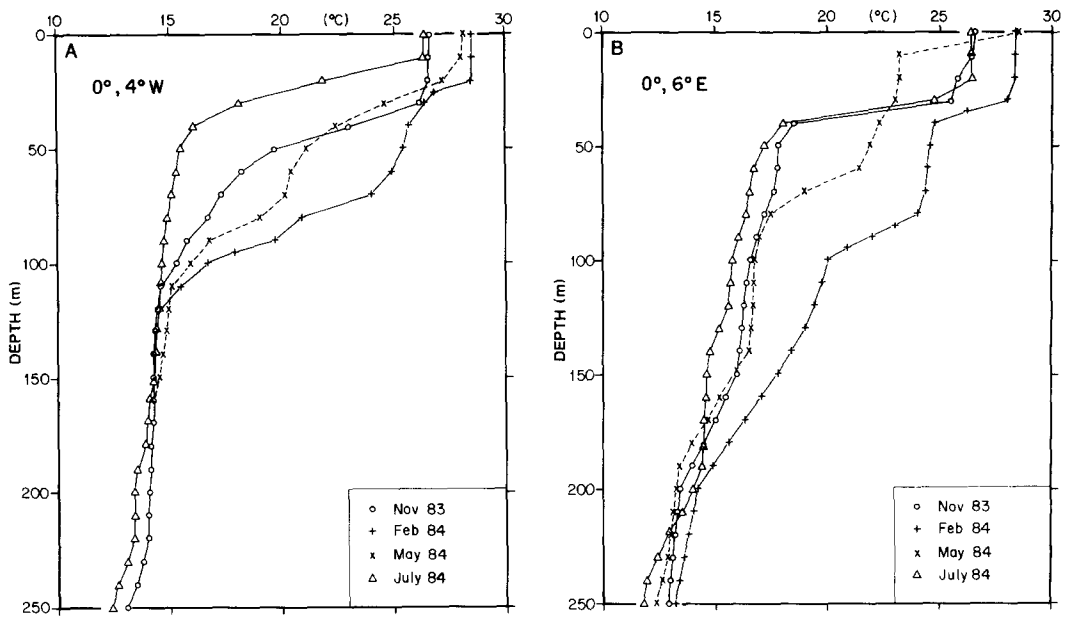


FIG.29. Equatorial temperature profiles (0-250m) from November 1983 to July 1984, at (a) 4°W and (b) 6°E. At 4°W, the downwelling-upwelling occurs within the upper 120m, whereas the signal is sensible at 250m at 6°E.

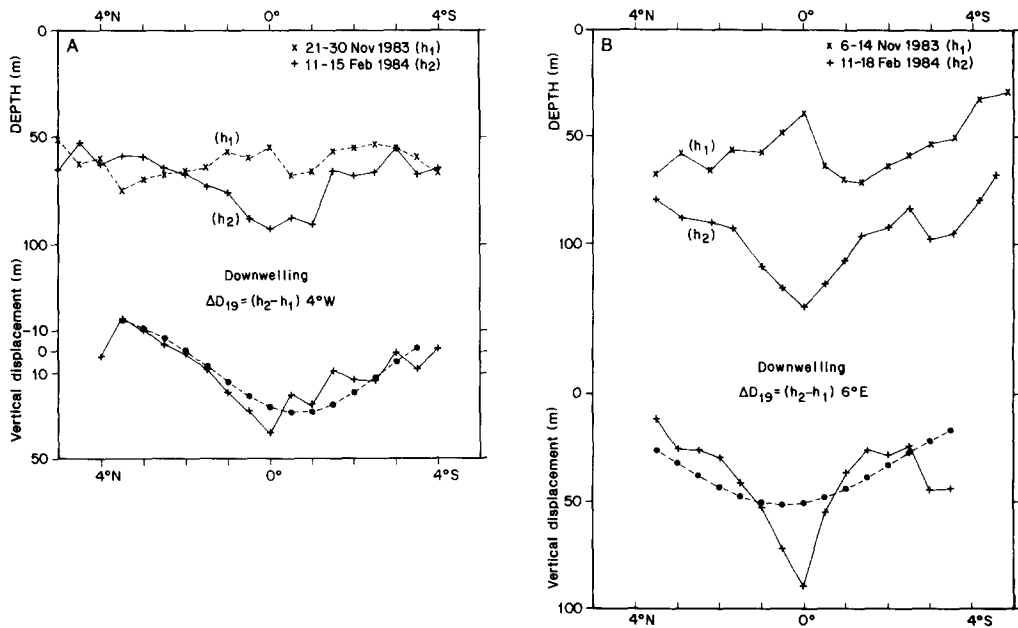


FIG.30. Depth of the 19°C isotherm at (a) 4°W and (b) 6°E during the downwelling situation from November 1983 (h1) to February 1984 (h2) and its vertical displacement Δd (19). The equatorial trapping is evident both at 4°W and 6°E. Each fit explained 88% and 47% of the variance, respectively.

During the upwelling onset (February - May 1984), the 19°C isotherm moved upwards (Fig.29a,b); the move was small at 4°W (only 13m) but very large at 6°E (62m). This change in depth was maximum at and decreased nearly symmetrically with the equator. A Gaussian fit at 6°E explained 65% of the variance with a trapping scale of 251km, a Kelvin wave speed of 144cm s^{-1} and an equivalent depth of 21cm. These scales are in good agreement with those associated with the second baroclinic mode in the equatorial Atlantic.

Finally, between May 1984 and July 1984, the upwelling was fully developed along the equator in the Gulf. At 4°W the 19°C isotherm was upwelled 52m at the equator but showed nearly no rise either at 2°30'N or 3°S (Fig.31a). The scales given by the fit are now 203km, 94cm s^{-1} and 9.1cm, all values near the second baroclinic mode. At 6°E, we observe a general uplift of about 40m in the 19°C isotherm between 3°N and 3°S, but no equatorial trapping at all (Fig.31b). It is possible that we are observing here the reflection of first meridional mode ($L=1$) Rossby waves against the continent.

Comparing the temperature profiles at 0°, 4°W and 0°, 6°E (Fig.29), we observe that the downwelling-upwelling phenomenon occurred within the upper 120m at 4°W, whereas the signal reached at least 250m at 6°E where a cooling of 1.4°C at 250m was observed between February and July 1984 (Fig.29b).

5.3.4 Changes in thermostat thickness. The thermostat thickness at 4°W (6°E) was defined as the layer comprised between the isanosteres 1.8ml kg^{-1} and 1.5ml kg^{-1} (1.6ml kg^{-1}). On the 1.8ml kg^{-1} isanosteric surface, as defined as the top of the thermostat, the temperature is near 16°C whereas the salinity increases from 35.63 at 4°W to 35.67 at 6°E (Fig.32a,b). There is a slight west-east change in the isanosteric surfaces that coincide with the bottom of the thermostat: at 4°W the thermostat is contained in the layer $1.5\text{--}1.8\text{ml kg}^{-1}$ while at 6°E it is contained in the layer $1.6\text{--}1.8\text{ml kg}^{-1}$. This change corresponds to higher temperatures and salinity at 6°E than at 4°W on the isanosteric surfaces coincident with the bottom of the thermostat (Fig.32a,b).

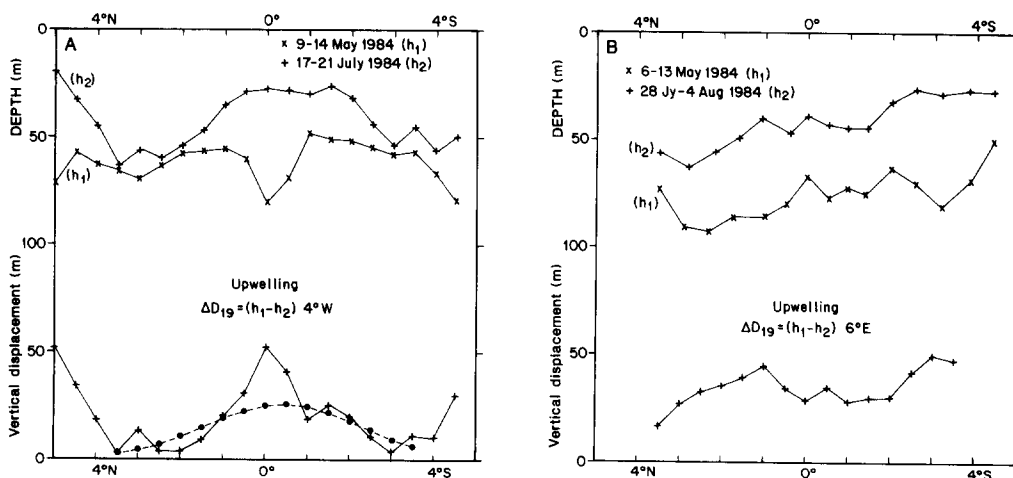


FIG.31. Depth of the 19°C isotherm at (a) 4°W and (b) at 6°E during the upwelling situation from May 1984 (h₁) to July-August 1984 (h₂) and its vertical displacement Δd_{19} . The equatorial trapping is evident at 4°W but has disappeared at 6°E. The Gaussian fit at 4°W explained 57% of the variance.

The thermocline thickness exhibits maxima 2° - 3° of latitude north and south of the equator and a minimum at the equator at 4° W as well as at 6° E (Figs 19d and 32a,b). The minimum at the equator is associated with a trough at the top and a ridge at the bottom of the layer. The equatorial depression represents the troughing of isotherms in the lower part of the thermocline, indicative of a high pressure zone and geostrophic balance for the EUC which flows usually above the 1.8 ml kg^{-1} isanosteric surface. The ridge at the bottom is a consequence of the convergence of water below the EUC, towards the equatorial low pressure zone.

From November 1983 to February 1984, the thickness of the thermocline at the equator decreased both at 4° W and 6° E, and increased at 2° - 4° N(S). In February 1984, the pinching of the thermocline between $1^{\circ}30'$ N and $1^{\circ}30'$ S is evident as well as its maximum vertical extension at 2° - 4° N,S (Fig.32). Deep eastward jets, separated from the shallow EUC core, appeared in February 1984 at about 300-400m at 4° W. Eastward flows were also observed below 200m at 1° E and 6° E. These deep jets tend to destroy the equatorial thermocline, as a result of the vertical spreading of the isotherms downwards below the EUC and upwards above 400m (Figs 28a, 32a, February 1984). It is noteworthy that the shallow core of the EUC did not change in depth during the downwelling event either at 4° W (60-70m) or at 6° E (60m), although its flow increased from about 6 to 7 Sverdrups at 4° W and from 2 to 3 Sverdrups at 6° E.

In May 1984, the thermocline thickness did not change at the equator but was decreasing at 2° N, 2° S.

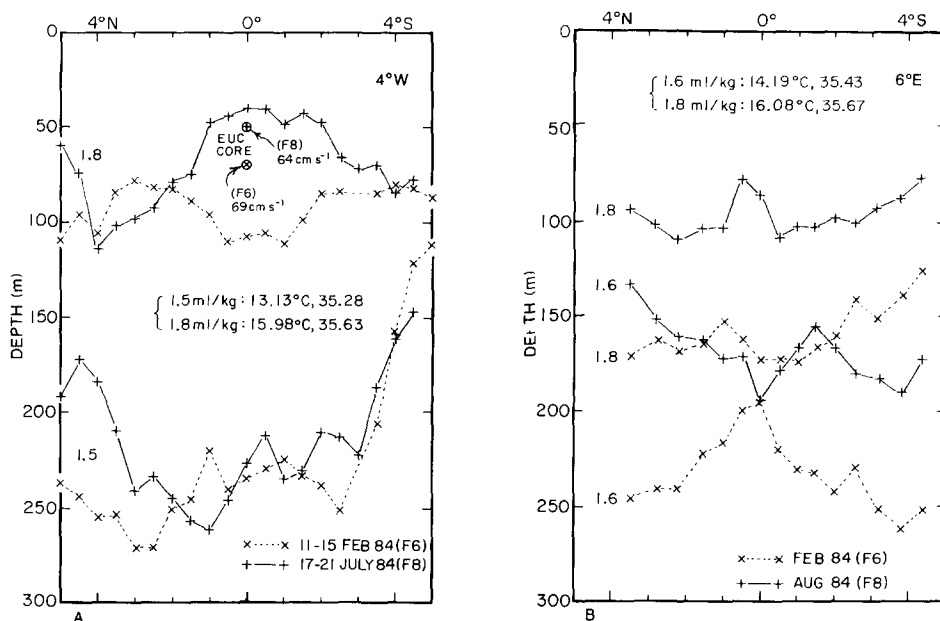


FIG.32. Thermocline thickness defined as the vertical distance between (top) 1.8 ml kg^{-1} and (bottom) 1.5 ml kg^{-1} (1.6 ml kg^{-1}) isanosteres at 4° W (6° E) between 5° N and 5° S. (a) At 4° W. Downwelling situation in February 1984, upwelling in July-August 1984. The core of the EUC is shown at 4° W (maximum speed). The depression (ridge) of the 1.8 isanostere at 1° N - 1° S in February 1984 (July 1984) implies higher (lower) pressure at the equator and divergence (convergence). At depth in February 1984, the equatorial ridge at 1° N - 1° S is surrounded by two depressions implying higher pressure and convergence. (b) At 6° E. The whole thermocline is uplifted by about 75m from February 1984 to August 1984. Note the depression (ridge) of the top of the thermocline at the equator implying divergence (convergence) in February 1984 (July 1984). On the horizon 1.6 ml kg^{-1} convergence in February 1984, divergence in July 1984.

Finally, in July-August 1984, the 1.8 ml kg^{-1} isanosteric surface had a ridge at the equator after an upward move of nearly 70-80m at 4°W and 6°E , and the thickness of the thermostad increased of 60m and 80m, respectively. Poleward of 2°N , the thermostad was strongly reduced (Figs 28b, 32). Poleward of 3°S , the thermostad thickness was nearly constant (Fig.32), indicating constant mixing conditions within the SEUC, although we have observed an acceleration of this flow at $4^\circ\text{-}5^\circ\text{S}$ (Fig.28).

5.4 Interpretation in terms of equatorial Kelvin waves

These analyses of the perturbation temperature fields along 4°W and 6°E show that the variations of the 19°C isotherm depth (Figs 30,31) and the changes in thickness of the equatorial thermostad (Fig.32) were strongly equatorially trapped. From February 1984 to July-August 1984, the changes in the thermostad thickness at 4°W (Fig.33) and 6°E (Fig.34) were maximum at the equator (increases of 60-90m at 4°W and 82m at 6°E) and nearly zero at 2°N , S. On the other hand, we do not observe a significant increase in speed or transport in the shallow core of the EUC during the downwelling as well as during the upwelling events. Rather, we observe an increase of the eastward zonal speed and transport at 300-400m at 4°W from November 1983 to February 1984 and May 1984. In May 1984, a distinct eastward deep jet appeared clearly at 250-350m with zonal speeds above 30 cm s^{-1} and an estimated transport of about 3 Sverdrups. In July 1984, this deep jet had shoaled to 250m, transporting about 4.6 Sverdrups (Fig.28b). These observations are not in contradiction with the expected effect of the downwards propagation of equatorially trapped Kelvin waves (McCREARY, 1981, 1984; GIESE and HARRISON, 1990): the EUC perturbations are

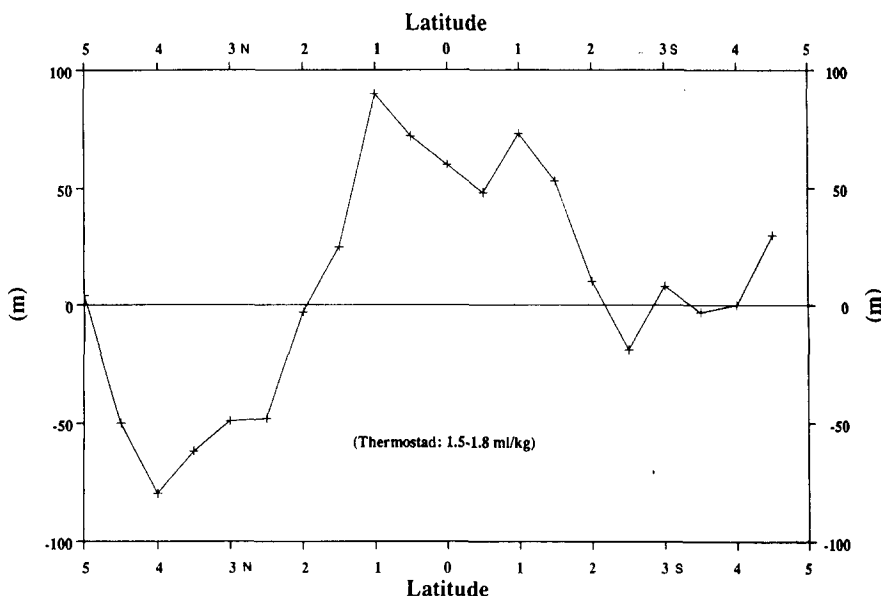


FIG.33. Changes in thermostad thickness at 4°W ($1.5\text{-}1.8 \text{ ml kg}^{-1}$) from February 1984 to July 1984, between 5°S and 5°N . Note the increase in thickness (positive values) between 2°N and 2°S ; the decrease (negative values) between 2°N and 5°N , concomitant with a deep eastward jet at the equator and a swift Guinea current, respectively. The deep eastward jet carries 4.6 Sverdrups and uplifts the top of the thermostad as well as the shallow equatorial thermocline.

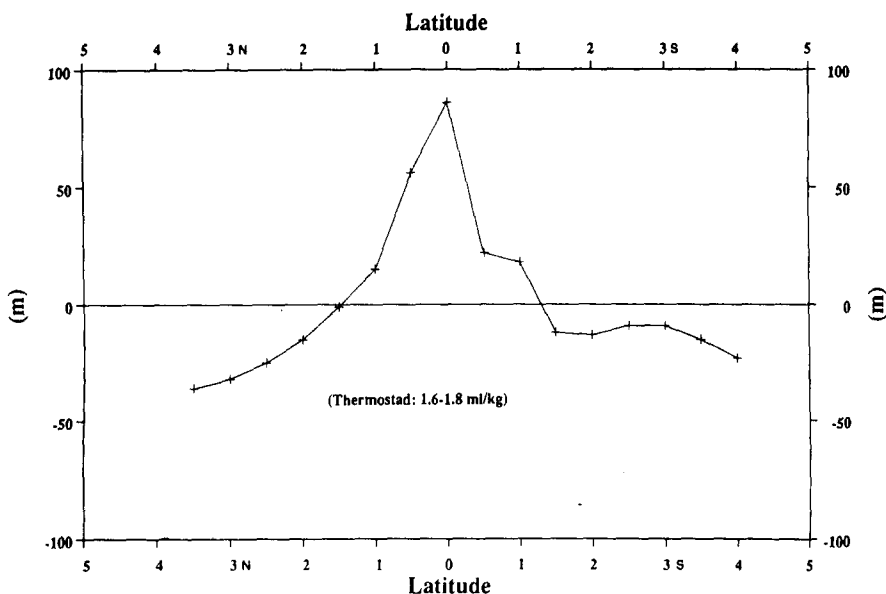


FIG.34. Changes in thermostat thickness at 6°E ($1.6-1.8 \text{ ml kg}^{-1}$) from February 1984 to August 1984, between 3°30'N and 4°S. Positive (negative) values indicate an increase (decrease) in thickness. Note the strong equatorial trapping between 1°30'N and 1°30'S.

seen first at depth (300-400m) and there is practically no change in the shallow EUC core at 50-70m.

Modelling studies (PHILANDER, 1979; McCREARY, 1981, 1984; PHILANDER and PACANOWSKI, 1981, 1986a,b; PHILANDER, 1990) and synoptic time series data at four locations along the equator during the FOCAL/SEQUENTIAL programmes provide evidence that the subsurface thermal structure and the sea surface slope adjust rapidly through long equatorial waves generated by the zonal wind stress variations. The evolution of the upper ocean equatorial response, based on synoptic observations during 1982-1984, has been described among others by KATZ (1987a), KATZ *et al.* (1986), WEISBERG and TANG (1987) and VERSTRAETE and VASSIE (1990). Comparing the ZPG time series at the 28°W and 4°W moorings with data derived from an equatorial long-wave analytic model forced by realistic zonal wind stress distribution, WEISBERG and TANG (1987) show that the ZPG adjusts very rapidly with the large-scale changes in τ_x . In their model, the reduced-gravity wave speed was fit with the phase speed of the second baroclinic mode used by BUSALACCHI and PICAUT (1983) and McCREARY, PICAUT and MOORE (1984). On the other hand, KATZ (1987b) found evidence of the first baroclinic Kelvin mode during 1983 and 1984 along the equator between 34°W and 1°W in the frequency band of 10^{-1} to 40^{-1} cpd. Studying the basin-wide zonal pressure gradient anomaly of February 1984 in the Gulf of Guinea, VERSTRAETE and VASSIE (1990) concluded that the

second baroclinic mode was also a constant feature during 1983-1984 in the central-eastern equatorial Atlantic (between St Peter and St Paul rocks and Sao Tomé island).

We have seen that the reversal of the ZPG between February 1984 and April-May 1984 in the Gulf of Guinea cannot be explained by local forcing. Moreover, there was no equilibrium between the vertically integrated ZPG (to 120m) and τ_x along the equator both in February 1984 and in April-May 1984. The depression-ridge of isotherms below the thermocline during the downwelling-upwelling event had a meridional structure that suggests that in February 1984, an equatorial downwelling Kelvin wave was involved, which was succeeded by an upwelling Kelvin wave in April-May 1984.

In consequence of the perturbations of the temperature (and density) fields, large dynamic height and ZPG anomalies were generated in February 1984 particularly at 6°E. It was impossible to relate these anomalies to the local forcing. VERSTRAETE and VASSIE (1990) gave evidence that the first empirical orthogonal functions (EOF) of the filtered sea levels at Principe (1° 39'N), Sao Tomé (0° 01'N) and Annobom (1° 25'S) had a meridional structure maximum at and symmetrical about the equator. A Gaussian fit of the EOF amplitudes was significant at the 0.01 level and the separation constant, c , was equal to the phase speed of the second baroclinic mode Kelvin wave in the Atlantic (140cms⁻¹).

On the other hand, a study of the cross-correlation of the detrended low-pass filtered sea level show that Sao Tomé was lagging RSPP by 33 days. This time lag also corresponds to the phase speed of the second baroclinic mode (VERSTRAETE and VASSIE, 1990).

A cross-correlation of 43435 hours (5 years) of coincident observations of the sea level and *in situ* temperature (at 5.6m depth) at Principe from November 1983 to November 1988 shows that the sea level precedes the near surface sea temperature by about 62 days. This is the time necessary for a cooling occurring at 500m to reach the mixed layer at the vertical phase speed of about 7.3m day⁻¹. This result is consistent with the seasonal cycle observed in the historical data at 6°E (2°N - 2°S) where we observe a cooling at 500m in April-May shifting to 50m in June-July (Fig. 14). This result is also in agreement with the upward phase propagation of the upwelling event off Abidjan at 7m day⁻¹ from 350-400m to the surface (Fig. 18) as observed by PICAUT (1983). According to McCREARY (1984), the energy that appears east of the wind forcing area is propagating eastwards and downwards at the slope

$$\text{tg}\alpha = -\sigma/N = c_z/c_x$$

for the Kelvin wave, where σ is the wave frequency, N the Vaisala frequency (supposed constant) and $c_z(c_x)$ is vertical (horizontal) phase speed. Between 0° and 30°W, the zonal wind stress oscillates at a frequency of about 6 months (Fig. 2a). Since the “downwelling-upwelling” signal occurred twice a year in the Gulf of Guinea, the appropriate frequency is $2\pi/6$ months. In an ocean of constant stratification N , it is possible to evaluate the slope $-\sigma/N$. From the observations below the tropical thermocline during 1982-1984, the mean Vaisala frequency was evaluated by integration between 100m and 500m. At 35°W (6°E) N was estimated to $5.4 \cdot 10^{-3} \text{ rad s}^{-1}$ ($4.3 \cdot 10^{-3} \text{ rad s}^{-1}$), resulting in a zonal slope of -74m/1000km at 35°W and -93m/1000km at 6°E.

Taking $c_z=7.3\text{m day}^{-1}$ (estimate of the vertical phase velocity at Principe in 1983-1988), $c_x=1.4\text{ms}^{-1}$ (1.2ms^{-1}) at 35°W (6°E), we find new evaluations of the slope: -59m/1000km at 35°W and -69m/1000km at 6°E, which are in reasonably good agreement with the values computed above from the background stratification N .

The semi annual signal in the zonal component of the wind stress appears clearly at 20°W - 30°W (Fig. 2a). Kelvin beams radiating from the ocean interior at say 25°W slopes downward into the eastern ocean and reaches about 300m at 6°E. At this very low frequency, vertical group velocity of such waves is so small (10m day⁻¹), that baroclinic modes are unlikely to be formed, since the waves will reach coasts before the ocean bottom. Since the zonal windstress in the equatorial strip 10°W - 35°W is able to equilibrate the integrated zonal pressure gradient in the upper 120m, it is able as well to generate Kelvin beams in the thermocline in this area. These beams are sloping more and more steeply downward as they progress eastward in the equatorial thermocline because N is reduced in this layer.

6. DISCUSSION AND CONCLUSION

The equatorial downwelling-upwelling in 1984 in the Gulf of Guinea has been shown to be associated with remote variations of the equatorial zonal wind field, but the connection between the equatorial and coastal upwellings is not obvious. PICAUT (1983) gives evidence of a southward propagation at 66cm s⁻¹ of the upwelling signal between 1°S and 13°S, and a westward propagation along the northern coast between 2°E and 8°W, at speeds varying from 36 to 90cm s⁻¹. However SST is not appropriate because it is influenced by advection and analysis confined to measurements along a coast are of little help as well. Studying the thermal structure at 4°W, HOUGHTON and COLIN (1986) found the same lag between the wind intensification at 29°W and the upwelling at 4°W (22 days for both 1983 and 1984), and this implies a phase speed of 1.3ms⁻¹ corresponding to the second baroclinic Kelvin wave. But the lag between the equator and the north coast was not constant (52 days in 1983 and 28 days in 1984).

The region between Sao Tomé, Annobom and Pointe-Noire (5°S) is of special interest since equatorial and coastal scales overlap there. Cross-correlations between coincident filtered sea level records in 1983, 1985-1986 and 1988 do not show a constant phase lag as expected. Similarly, we have failed to find constant phase lags between Lomé, Takoradi and Abidjan. Although Lomé leads the two other stations, the phase speeds vary between 0.34 and 1.0ms⁻¹. The shoaling of the thermocline in 1984 at 4°W (Fig. 31a) had a width scale of about 180km corresponding to the local Rossby radius of deformation at 5°N for the first baroclinic mode. Unfortunately, there was no coastal trapping during 1983. These results are conflicting with a coastal upwelling forced by a coastal trapped wave originating at the equator and propagating polewards along the eastern boundary (the remote forcing hypothesis).

The remote forcing hypothesis was evaluated by PHILANDER and PACANOWSKI (1986a) in a simulation with a general circulation model (GCM) forced with climatological winds. In this experiment, a wall along 35°W replaced the coast of South America, and the winds along the equator were only westward (neither eastward nor meridional winds and no curl). It was found that the generated eastward coastal current (the Guinea current) and the upwelling near the coast (5°N) were very small. This means that the westward winds alone to the west of 0°W along the equator have only a minor effect on the coastal upwelling at 5°N. According to other numerical experiments, PHILANDER and PACANOWSKI (1986a) showed that the northward component τ_y and the curl of the wind to the north of the equator are more important in determining the seasonal changes at the coast. On the other hand, the shoaling of the thermocline over the Gulf of Guinea from May onwards, make it possible for the winds parallel to the coast to induce coastal upwellings. At 5°N, the winds have an eastward component and cause the thermocline to slope downward to the east. Along the Ivoirian-Ghanaian shelf (5°W - 5°E), the eastward winds (Fig. 2b) and this slope (Fig. 21a) are more intense in July-September. The westward pressure force

associated with this slope drives a westward coastal undercurrent, the Guinea coastal undercurrent observed by LEMASSON and RÉBERT (1973a,b). In the area $2^{\circ}\text{N} - 5^{\circ}\text{S}$ and to the east of the first meridian, the winds are essentially northward (Fig.2b,c area $0^{\circ} - 10^{\circ}\text{E}$) and the sea surface slope is southward (Fig.21a). This slope is associated with southward pressure gradients (Fig.21b,c) and a poleward undercurrent, which was observed along the Gabon-Congo shelves (WACONGNE, 1988).

The analysis of the thermocline water masses suggest that the EUC supplies the coastal upwellings because the Guinea and the Gabon-Congo undercurrents carry a well identified South Atlantic central water type, nearly identical to the SACW carried by the EUC. This implies a northward and southward extension of the EUC. The observed splitting of the EUC core (Fig.12b) in 1982-1984 at Sao Tomé island lying at $0^{\circ} - 0^{\circ}30'\text{N}$ is consistent with the geostrophic westward flows observed in the thermocline to the north and south of the EUC. These observed westward compensation flows, flanking the EUC at $2^{\circ} - 3^{\circ}\text{N}$ and S, coincide with the observed maximum thickness of the thermocline at these latitudes.

The coincident records of sea level and superficial sea temperature at 6°E show that the sea level changes precede those in the near sea-surface temperature by about one to two months and this is an indication that the deep signal leads the surface signal in the Gulf of Guinea. The seasonal fluctuations of the thermal and salinity fields are observed to depths as great as 500m over the Gulf.

In 1982-1984, in the western-central equatorial Atlantic, the ocean was in equilibrium with the winds, save during the period December - May 1984 when the trade winds collapsed. In the Gulf of Guinea, the ocean was not in equilibrium with the wind. There, the distinct semi-annual cycle is attributable to the semi-annual signal in the zonal winds to the east of 30°W (Fig.2a), not to the winds off South America (PHILANDER and PACANOWSKI, 1986a). The downwelling-upwelling events in the eastern equatorial Atlantic are excited by abrupt relaxation-intensification of the trade winds in the central-western equatorial Atlantic and can be explained by the equatorial oceanic adjustment to this wind forcing. The abrupt relaxation of the TW in December 1983 to January 1984 as well as the impulsive intensification of the trade winds in April-May 1984 are the essential features of the equatorial remotely wind forced downwelling-upwelling in the Gulf during 1984.

Exceptionally, the abrupt relaxation of the trade winds in December 1983 - January 1984 occurred over the whole basin. The collapse which followed from January to April-May 1984 had a duration nearly equal to the intrinsic adjustment time of the equatorial Atlantic. According to CANE (1979) and PHILANDER (1981a), this time is estimated to be 150 days. At these space and time scales, the Kelvin waves have wavelengths nearly equal to the size of the basin, and they generate basin-wide zonal pressure gradients as observed. The pressure and zonal velocity perturbation fields could strongly penetrate below the shallow thermocline EUC flow. The facts that at 4°W and within $1^{\circ}30'\text{N} - 1^{\circ}30'\text{S}$ (1) the observed westward flow at 400m in November 1983 reversed to an eastward flow in February 1984; (2) the eastward flow was increasing significantly at depth and not in the shallow EUC core from February 1984 to July 1984; (3) that the observed deep eastward jet, isolated from the EUC core, increased its flow from 1 to 4.6 Sverdrups and rose from 325m in February 1984 to 250m in July 1984 are in agreement with the theory of a remotely forced deep equatorial jet generated by equatorial Kelvin beams (McCREARY, 1984; GIESE and HARRISON, 1990). Only continuously stratified or GCM models are able to describe this penetration of the signal at depth, below the shallow thermocline.

This work indicates that subsurface observations of temperature and zonal velocity in the thermocline region, combined with sea level observations, are very useful in the identification of the remotely forced Kelvin response in the eastern equatorial Atlantic. Until now, efforts to detect

these waves in the real ocean have been handicapped by the paucity of subsurface data (CHAO and PHILANDER, 1991). The thermal and sea level perturbation fields either at 4°W or 6°E were strongly trapped at the equator, and their scalings give evidence of a second Kelvin baroclinic beam in the equatorial Atlantic. Since the equatorially trapped waves are able to explain some aspects either of the 1984 winter warm event or of the annual cycle, it is very likely that the annual cycle as well as the “quasi El Niño” or the warm events in the eastern equatorial Atlantic are driven by the same dynamics. The 1984 warm event in the equatorial Atlantic was phased-locked to the annual cycle and this is also characteristic of the ENSO events in the equatorial Pacific (WYRTKI, 1974; MEYERS, WHITE and HASUNUMA, 1982).

The coastal upwellings in the Gulf of Guinea have a considerable latitudinal scale and are not a local coastal phenomenon. Any further speculation about the relationship between the equatorial and coastal upwellings in the Gulf of Guinea would require reliable synoptical and deep-sea observations reaching at least the Antarctic Intermediate Waters all along the eastern boundary as well as at the equator. Historical data as well as the very few simultaneous tide-gauge stations do not allow us to address now the important questions of the poleward propagation of the seasonal upwelling along the N-S boundary and the reflexion of the impinging Kelvin waves.

Finally, this study suggests that increasing the vertical resolution of the GCM numerical models below 150m is necessary in order to describe the deep equatorial jets as well as the subthermoclinical compensation flows such as those observed at 2°-3°N and 2°-3°S in the Gulf of Guinea. Such essential features cannot be simulated by shallow-water models and are not adequately modelled by GCMs now. This situation detracts from our ability to assess properly the potentials of the TOGA/WOCE numerical experiments in the equatorial Atlantic Ocean.

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